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PROGRAM MANAGER; Laura J. Kelland-May.
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THE EVALUATION AND MEASUREMENT OF EVAPORATION, TRANSPIRATION, INTERCEPTION, AND EVAPOTRANSPIRATION

by

MICHAEL G. MIHAN

Abstract

Five components of the hydrologic cycle: interception, evaporation, transpiration, evapotranspiration, and infiltration, are reviewed. The nature and process of each component, as well as, the factors that influence each component, are discussed. The various methods of measuring and estimating each component are presented.

A section on evaporation networks is included. A discussion of optimum and minimum networks, with emphasis on New Brunswick, is provided.

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LIST OF SYMBOLS

AE	-	Actual evapotranspiration
A_f	-	Area of forested land
a	-	Coefficient
B	-	Mean barometric pressure
D_c	-	Mean canopy density
D_A	-	Depth of water on the soil surface
E	-	Evaporation
e	-	Base of natural logarithms
E_a	-	Drying power of the air
E_{at}	-	Evaporation rate from open water
e_a	-	Vapor pressure of the air
e_{mi-2}	-	Saturation vapor pressure corresponding to average daily minimum temperature minus 2°C
e_{ma}	-	Saturation vapor pressure corresponding to maximum daily temperature
e_1	-	Vapor pressure at lower height above the surface
e_2	-	Vapor pressure at upper height above the surface
e_s	-	Saturation vapor pressure at the water surface temperature
F	-	Depth of water infiltrated
F_i	-	Stored ice
f_c	-	Infiltration rate at cessation of rainfall
f_r	-	Infiltration rate
HI	-	Annual heat index
H	-	Heat budget
h	-	Height
I	-	total interception loss

I_c - Interception loss for a closed canopy forest

I% - Percentage interception

i - Rainfall intensity

I_B - Interception loss for a basin

I_f - Effective heat

In - Inflow

k - Coefficient

L - Latent heat of evaporation

M_p - Monthly percent of annual daytime sunshine

n/N - Ratio of actual to possible hours of hours sunshine

O - Outflow

Og - Seepage

P - Gross precipitation

P_a - Annual precipitation

PE - Potential evapotranspiration

p - Density of water

Q_c - Discharge rate of cessation of rainfall

Q_r - Measured discharge rate at any time during the cecession

R - Ratio of vegetal surface area to its actual area of canopy

RH - Mean relative humidity

r - Radiation reflection coefficient

R_A - Theoretical maximum solar radiation that reaches a site in the absence of atmosphere

R_e - Energy utilized for evaporation

R_l - Net longwave radiation exchange between the atmosphere and ground

R_o - Increase in energy storage

R_r - Reflected solar radiation

R_s - Sensible heat loss from a body of water

R_t - Solar radiation incident to a surface

R_v - Net energy advected into a body of water

S - Storage capacity of vegetation

S_d - Depression storage

S_f - Stemflow of water

SR - total depth of runoff

T - Mean air temperature

T_a - Actual air temperature

t - Duration of rainfall

T_h - Throughfall

T_w - Temperature of water surface

s - Change in storage

W - Wind speed

W_1 - Wind speed at lower height above the surface

W_2 - Wind speed at upper height above the surface

$\triangle$ - Slope of saturation vapor pressure curve at mean air temperature

γ - Constant

σT_a^4 - Theoretical black body radiation at mean air temperature

CHAPTER 1
INTRODUCTION

The chief practical significance of the hydrologic cycle lies in the fact that it is the process that supplies water to plants and animals, and to man for his consumption. The determination of the components in the hydrologoic cycle must be given careful consideration before the available water resources of a drainage basin can be satisfactorily evaluated.

A thorough understanding of interception, evaporation, transpiration, evapotranspiration and infiltration, is essential for hydrologists or hydrometeorologists seeking to determine the water balance and potential water supply of a drainage basin. In addition, the hydrologist must also be able to deal quantitatively with the interrelations between the components and be familar with the various techniques and instruments available for measurement.

The primary objective of this report is to summarize an extensive literature review of interception, evaporation, transpiration, evapotranspiration and infiltration. The nature and factors that influence each component are discussed in detail.

The secondary objective of the report is to identify instruments and empirical equations used to measure each component with consideration of their application to New Brunswick.

Each of the components are discussed in separate chapters, followed by a chapter dealing with evaporation networks. This chapter has emphasis on New Brunswick needs, and is included to illustrate the overall collection of evaporation data.

Interception, evapotranspiration, and infiltration are often neglected or overlooked when dealing with water resource assessments. However, it is hoped that assumptions about these components can eventually be replaced by measured data using the means identified in this report.

CHAPTER 2
INTERCEPTION

2.1 GENERAL

A portion of the precipitation falling to the earth's surface may be stored or collected by vegetal cover and subsequently evaporated. That volume of water caught by vegetation is referred to as interception, and the portion thereof retained and evaporated is called the interception loss (Gray, 1970).

Interception may be classified as litter interception or canopy interception. Litter interception is the rainfall retained on the forest litter layer and evaporated without adding to the moisture in the underlying mineral soil (Zinke, 1965). Litter interception is often neglected because it only comprises a small amount of the total interception.

Canopy interception has been defined as that portion of precipitation retained on standing vegetation and evaporated (Zinke, 1965). Canopy interception is important hydrologicly, because it causes non-uniform wetting of forest soils, inhibits transpiration and adds significantly to total vaporization losses (Lee, 1980).

The modifying effect which a forest canopy can have on rainfall intensity at the ground can be put to practical use in watershed management schemes (Viessman, 1972).

Interception should not be neglected, because, depending on the nature of the cover and precipitation characteristics, it may be a very significant factor in water balance studies (Gray, 1970).

2.2 PROCESS OF INTERCEPTION

The gross precipitation reaching an area covered by vegetation is divided into net rainfall and interception. Net rainfall is the quantity of water that actually reaches the forest floor. It is the sum of throughfall and stemflow.

Throughfall is that part of the gross precipitation that reaches the ground directly through the vegetative canopy, through intershrub spaces in the canopy, and as drip from leaves and twigs (Chow, 1964). The largest proportion of net precipitation reaching the forest floor is throughfall.

Stemflow is that part of the gross precipitation that reaches the ground by running down the stems of trees (Chow, 1964). Stemflow starts only after a certain amount of rain has fallen, and has been found to increase with the amount of precipitation per shower when the rainfall is not a limiting factor (Kittredge, 1948).

The gross precipitation retained in the canopy, on the surface of leaves, twigs, and stems, is partly absorbed by the vegetation or slowly begins to spread into a thin layer to collect at edges, ridges and depressions (Gray, 1970). This process continues until the canopy storage capacity is

reached, and surface tension forces are in balance with gravitational forces. Canopy storage capacity is defined as the quantity of water that can be held on the above ground portions of a forest stand (Zinke, 1965). It is expressed in the same units as precipitation, as an average depth for the stand in terms of liquid water eqvivalent (Lee, 1980). The canopy storage capacity is dependent on:

1) position of leaf surface (horizontal or inclined);
2) the form of the leaf;
3) surface tension relations between the water and the leaf surface;
4) wind velocity;
5) intensity of rainfall; and
6) size and impact of falling drops (Gray, 1970).

Once the storage capacity is reached, further rain striking the canopy will cause drops to separate from the edges of the leaves and fall down to strike a lower leaf or pass through to the ground.

Evaporation losses also occur from the surface of the canopy and must be considered as interception loss. The rate of evaporation increases until storage capacity is reached and then it remains constant (Linsley, 1975).

The total interception loss does not represent a corresponding loss of available water in the soil. The loss of water by interception is compensated to some extent during the growing season by a reduction in transpiration (Leyland

et. al., 1965). However this reduction is relatively small in comparison with the amount of interception loss. Sykes (1960) and Wells (1963), working with potted tree seedlings or detached shoots and leaves, have shown that the rate of evaporation of intercepted water can be in the order of five to ten times that of transpiration (Leyland et. al., 1965). This indicates that although wetting the foliage reduced both water uptake and transpiration, the net interception loss is over 90 percent of the total interception.

2.3 INTERCEPTION OF SNOW

Snow reaches the canopy and is stored somewhat differently than raindrops and therefore differences occur in the amounts of interception. The interception of snow by forests is a major factor in flood runoff predictions and therefore is of great importance in areas where flooding is a problem (Gray, 1970).

In forests snow particles move almost horizontally at low speeds and interception is increased due to a greater surface area of canopy available. Snow interception is negligible in hardwood stands because of a lack of foliage and the surface area for interception is drastically reduced (Kittredge, 1948).

The interception loss may only be a small percentage of the total snow intercepted. Evaporation and sublimation from snow requires considerable more energy than does melting and

therefore most of the intercepted snow will reach the ground as drip after melting (Gray, 1970).

2.4 ANALYSIS OF INTERCEPTION

In order to do a complete review of interception, some basic equations which explain the process must be discussed. The equations review the factors involved in calculating interception loss, percentage of interception loss, and basin interception. An interpretation of the data, including the transferability of the data is also given. Various values of interception obtained from studies, and the transferability of these values from location to location is analized.

2.4.1 Interception Equations

Total interception loss is made up of the sum of two components: 1) water stored on the vegetation at the end of a storm; and 2) evaporation from wet plant surfaces during the storm (Gray, 1970). A mathematical relationship exists between these two components and can be expressed as:

$$I = S(1-e^{-P/S}) + REt \qquad \text{(Eq 2-1)}$$

I = total interception loss for the projected area of canopy;

S = storage capacity of vegetation for the projected area of canopy;

R = ratio of vegetal surface area to its projected area of canopy;

E = evaporation rate from the projected area of canopy;

t = duration of rainfall;

P = gross precipitation; and

e = base of natural logarithms

This equation indicates an exponential increase in the storage as the amount of precipitation increases. This trend has been observed in practice (Gray, 1970). The equation also shows that when precipitation is greater than storage capacity, the term $(1-e^{-P/S})$ approaches a value of unity. Storage capacity and evaporation are the only factors involved in interception when this is the case.

To obtain an estimate of the percentage of interception loss, the amount of interception is divided by the gross rainfall. Gross rainfall is calculated by multiplying the intensity of the rain by the duration. Interception expressed as a percentage is given by the equation:

$$I\ \% = 100\ \frac{[S(1-e^{-P/S}) + REt]}{it} \qquad \text{(Eq 2-2)}$$

where numerator = as defined above;
i = rainfall intensity; and
t = rainfall duration.

This equation shows two things. First, when the intensity of rainfall is kept constant, the percentage interception loss will decrease as the duration of the storm

increases. Second, the percentage interception decreases when the intensity of rainfall increases if the duration of the rainfall is kept constant.

When a watershed basin is composed of simple forests and is relatively open the following equation may be used to estimate interception (Zinke, 1965).

$$I_B = A_f - D_c - I_c \qquad \text{(Eq. 2-3)}$$

where: I_B = interception loss from the basin;
A_f = area of forested land;
D_c = mean canopy density; and
I_c = interception loss for closed canopy of the forest type concerned.

When the watershed basin is composed of many vegetation types, such as that found in New Brunswick, the basin interception is equal to the sum of the individual losses for each forest and vegetation type multiplied by the proportionate area of each type.

2.4.2 Interpretation of Interception Data

Numerous data have been collected by investigators concerning magnitudes of interception (Delfs, 1965; Zinke, 1965; Gray, 1970). Unfortunately, most of the values are only of purely local significance since interception is a function of the local rainfall, climate, and vegetation

(Leyland et.al., 1965). For a more general application of the data it is necessary to distinguish climatic influences from vegetative influences which as of today has not been accomplished.

A study in Maine from 1965-1966 has determined interception values for spruce-fir and birch stands (Table 2.I).

Table 2.I. Weekly Average Precipitation Catch of Standard U-S. Weather Bureau-Type Rain Gauges Located in Spruce-fir Stand and Hardwoood Stand During 1965-66 (Viessman, 1972).

Weekly Avg. Precip. Catch		Percent Interception	
Spruce-Fir	Birch	Spruce-Fir	Brich
0.24	0.33	38	15
1.01	1.25	30	14
1.01	1.23	26	10
1.41	1.65	21	8
0.55	0.81	37	7
0.66	0.95	39	12
0.20	0.25	23	4
0.36	0.55	41	10
0.25	0.58	58	2

Since The State of Maine has a similar climate and vegetation type as that of New Brunswick the values obtained in Maine would be comparable to estimates obtained in New Brunswick.

2.5 INTERCEPTION BY FORESTS

In New Brunswick the predominant vegetative cover is forest and therefore a discussion of the factors which affect

forest interception is given. The major factors that influence forest interception include: season; rainfall intensity; stand age; species; site quality; and wind.

2.5.1 Season

The amount of interception, for all species of trees, is greatly influenced by the season of the year. Interception reaches a maximum during the summer and a minimum during the winter. The difference between the amount of interception from summer to winter may be as much as 10-30 percent for most species (Kittredge, 1948).

There are two reasons for the reduction in the winter months: decreased canopy storage capacity and decreased amount of evaporation. The canopy storage capacity is reduced, for both coniferous and deciduous species, because of a loss of foliage from the trees. Evaporation from the tree surface is reduced during the winter months because the saturation deficit of the air is normally smaller at this time of year.

2.5.2 Rainfall Intensity

Rainfall intensity may have a large effect on the amount of interception. When the rain is light, of short duration or frequently interrupted the percentage of interception will be large. When rain falls in heavy showers or is persistent over a wide area the percentage of interception will be

small. In a mixed forest, 70.5 mm rain in 50 hours gives 25% interception, whereas a thundershower of 74.6 mm in 3.5 hours only gives 2% interception (Geiger, 1980). The difference in the interception is due to the force with which the rain strikes the canopy. High intensity storms cause large amounts of spattering, thereby reducing the amount of water adhering to the leaf surfaces.

2.5.3 Stand Age

Interception increases with stand age until stand maturity and then decreases as stands get older and begin to break up. For spruce approximately 10% of gross precipitation is intercepted at age 15, 35 % intercepted at age 80 and 20% at age 100 (Delfs, 1965).
For beech in Switzerland, it has been shown that the percentage of interception increases from two to twenty-seven percent from ages twenty to fifty, and then decreases to seventeen percent at age 90. (Table 2.II).

Table 2.II Variations in mean annual interception for beech in Switzerland (Kittredge, 1948).

Age Years	20	50	60	90
Mean Annual Interception (%)	2	27	23	17

The increase in interception with stand age until stand maturity is chiefly the result of increasing crown length and

denser foliage as stands get older. This results in a greater canopy storage capacity with age, and therefore, an increase in interception. The decrease after maturity is a result of the stands breaking up. This results in a loss of foliage and a decrease in canopy storage capacity.

2.5.4 Species

The amount of interception is influenced to some extent by the species of tree. The difference in interception is particularly noticable between coniferous and deciduous trees. Coniferous species generally intercept more rainfall than deciduous species (Kittredge, 1948). The lower values for deciduous species are the result of smaller crown storage capacities and extremely small amount of moisture required for wetting stems and branches (Kittredge, 1948). A study by the U.S. Forest Service in North Carolina support this observation (Figure 2.1).

The percentage interception also increases as species progress toward a more advanced stage of natural succession. Jack-pine, an intolerant pioneer species, intercepts approximately 20%; whitepine, mid successional species, intercepts 35% and hemlock, tolerant climax species, intercepts 50% (Kittredge, 1948).

The increase in interception with successional stage is due to an increase in canopy storage capacity with successional stage.

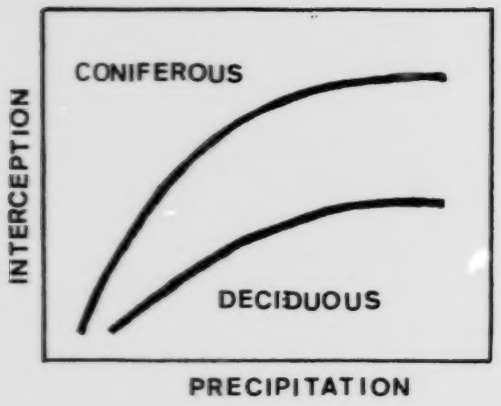

Figure 2.1 Interception in coniferous and deciduous forests as a function of precipitation (Wisler, 1978).

2.5.5 Site Quality

A very great influence on interception due to site quality can be assumed. It has been shown that the weight of foliage produced increases with site quality class for all species of trees (Delfs, 1965). Consequently interception increases as site quality improves due to an increase in canopy storage capacity.

2.5.6. Wind And Fog

The direction and speed of wind within a forest will have a large effect on the amount of interception. If there is wind accompanying the rain, then, owing to the motion of the leaves and branches, it is possible that maximum canopy storage capacity is materially reduced as compared with still air (Gray, 1970). This reduced storage capacity results in

decreased interception. However, wind may also increase interception loss by increasing the rate of evaporation from the canopy. In general wind will increase the total interception loss for a long storm and decrease it for a short storm (Gray, 1970).

The effect of fog is to decrease the total interception loss (Zinke, 1965). Moisture from the fog condenses on the leaves in the canopy thereby reducing the amount of rainfall required to reach maximum canopy storage capacity. This has the effect of decreasing interception.

2.6 INTERCEPTION FROM AGRICULTURAL CROPS

Not much information is available concerning interception by agricultural crops. The total amount of interception varies with spacing, age, size of plant, and amount of rainfall (Table 2.III).

Table 2.III. Interception by agricultural crops during different months of the year (Penaman, 1953).

		INTERCEPTION %		
		July	Aug	Sept
Rain	(mm)	86	53	0.6
Corn	(9/m^2)	4	22	100
	(36/m^2)	8	32	100
Oats	(16/m^2)	4	32	14
	(49/m^2)	10	36	29
Beans	(24/m^2)	27	22	57
Peas	(25/m^2)	-1	--	--

Values for interception by agricultural crops are extremely variable. This suggests that the only way to obtain accurate values of interception are through specific local field studies.

2.7 MEASUREMENT OF INTERCEPTION

Direct measurement of interception is virtually impossible because of the many variables in the process. Measurements therefore, usually involve a comparison between the amount of precipitation reaching the forest floor and the amount of gross precipitation. In order to make this comparison, gross precipitation, throughfall, and stemflow must be measured. These factors are related by the equation:

$$I = P - T_h - S_f \qquad \text{Eq. (2-4)}$$

where: I = interception loss;
P = gross precipitation;
T_h = throughfall of water; and
S_f = stem flow.

Each of the terms on the right hand side of the equation must be evaluated separately to find a value of interception.

2.7.1 Gross Precipitation Measurement

The measurement of precipitation available over the vegetal cover is normally accomplished using standard rain gauges and standard operating procedures (Gray, 1970). The sites chosen for placement of the gauges should be in large open areas near the interception study plots. Study plots

are normally no more than 2 to 3 hectares large.

The unavailability of large open sites near the interception study plot is a common problem. To overcome this problem, small clearings may be used provided there is an unobstructed sky view at an angle of 45 degrees from the horozontal surrounding the gauge opening (Gray, 1970). If no large openings or small clearings are available, gauges should be placed at tree top level to measure gross precipitation.

2.7.2 Throughfall Measurement

The method used to measure throughfall is normally determined by the type of vegetal cover present. For low bushes or shrubs simple sheet-metal troughs are frequently used as gauges. Trough gauges must be positioned carefully to account for variations in precipitation under the canopy and give a reading representative of the area (Kittredge, 1948).

If the vegetal cover is tall than standard rain gauges may be used to measure throughfall. Because of the effects of variation in throughfall, a large number of rain gauges are needed to get a representative value. Through the work with mixed forests in eastern U.S., the number of gauges necessary to sample throughfall to a 5% standard error of estimate as a function of the amount of throughfall, was determined (Table 2.IV).

Table 2.IV. Number of gauges necessary to sample throughfall to a 5% SE as a function of throughfall (Zinke, 1965)

Throughfall (inches)	# of Gauges Dorment Season	Growing Season
< 0.19	24	48
0.20 - 0.39	6	18
0.40 - 0.59	6	14
> 0.60	6	13

One other method used to determine the number of rain gauges necessary to sample throughfall, is to desiginate plots in the study area representing various conditions. Within each plot one rain gauge is placed at a random location.

2.7.3 Stemflow Measurement

It is practically impossible to measure stemflow from vegetation with numerous small stems. However, studies have indicated that stemflow is only a small proportion of interception in such cover and can therefore be neglected (Gray, 1970).

The measurement of stemflow for forest cover has been carried out using collars incised in the bark of the trees. Water flowing down the stem enters the collar through a pipe to collect in containers (Leyland et. al., 1965).

Stemflow may range from 1 to 15 percent for smooth bark species and from 2 to 3 percent for some rough bark species

(Chow, 1964). It has been shown that deciduous trees have a greater stemflow than coniferous trees (figure 2.2).

Since stemflow estimates may be very small they should be collected from several trees in a plot to give good estimates of average conditions. Helvey and Patric (1965) indicate that several plots are required to obtain accurate results, and they suggest that 5 to 10 randomly located plots should provide reliable results (Gray, 1970).

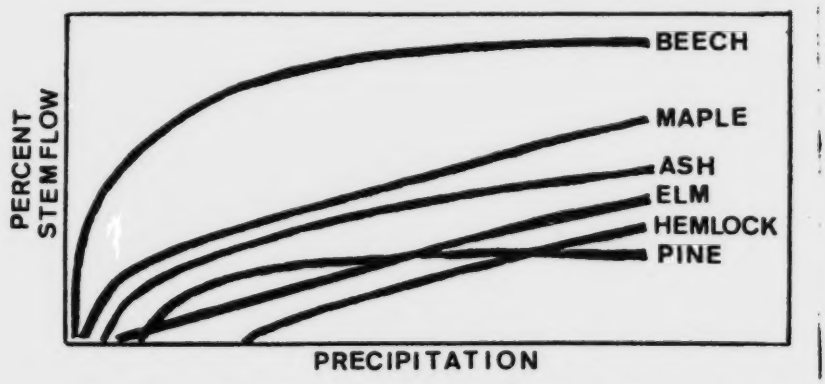

Figure 2.2 Percentage of stemflow for different species in relation to precipitation per shower (Kittredge, 1948).

CHAPTER 3
EVAPORATION

3.1 GENERAL

Evaporation is one of the principal components of the hydrologic cycle and is of great importance in many scientific fields related to water management. Water losses due to evaporation, in many cases, may be significantly large and, therefore, must be considered. Evaporation must be carefully evaluated to adequately establish water balances for lakes and reservoirs.

3.2 PROCESS OF EVAPORATION

Evaporation is the change of state of water from a liquid to a gas through vaporization (Meinzer, 1948). Vaporization occurs when water molecules attain enough kinetic energy to eject themselves from the water surface. Similarly some of the water molecules in the atmosphere may penetrate the water surface and remain in the liquid. Evaporation occurs if more molecules are leaving the water surface than are returning. The net exchange of water molecules per unit time at the liquid surface determines the rate of evaporation (Lee, 1980). Further evaporation can take place only when there is a supply of energy to provide the latent heat of evaporation (540 cal/g H20 @ 100 C) (Gray, 1970).

3.3 FACTORS AFFECTING EVAPORATION

The rate of evaporation is both a function of the climatic conditions and the characteristics of the evaporating surface. Climatic factors include: vapor and atmospheric pressure; temperature; wind speed; and radiation. Characteristics of the evaporating surface which influence evaporation include: the amount of soluable solids; size; shape; and depth. These factors are all interdependent and are difficult to determine singularly because they act in concert to determine the rate of evaporation.

3.3.1 Vapor and Atmospheric Pressure

The difference in vapor pressure between an evaporating surface and the surrounding atmosphere is a basic determinant of the evaporation rate. If all other factors remain constant, evaporation is proportional to the deficit in vapor pressure (Lee, 1980). The deficit in vapor pressure is the difference between the pressure of saturated vapor at the temperature of water and the vapor pressure of the air (Gray, 1970). As the vapor pressure deficit is increased the rate of evaporation is increased.

A decrease in barometric pressure with increased altitude results in an increase in the rate of escaping water molecules from a free water surface because there are fewer molecules in the atmosphere above the evaporating surface

(Chow, 1964). However, decreasing evaporation with increasing altitude would occur only if all other climatic factors remained the same.

3.3.2 Temperature

Evaporation is affected by both the temperature of the air and the temperature of the water. As the water temperature is increased the kinetic energy of the water molecules is increased causing the rate of vapor emission to be greater (Hounam, 1973).

The vapor holding capacity of the air varies with temperature. Increasing air temperature results in an increase in the vapor pressure deficit between the water and the air (Lee, 1980). This increase in the vapor pressure deficit leads to an increase in the evaporation rate.

3.3.3 Wind Speed

Varying wind speeds cause changes in the vapor pressure deficit and therefore cause changes in the evaporation rate. Wind movement replaces the air in contact with the water, by air which may have a different temperature or moisture content (Hounam, 1973).

The nature of the wind movement may also affect evaporation rates. Turbulent wind movements will remove water vapor from contact with the water surface and increase evaporation (Meinzer, 1948). Laminar movement may not materially

affect evaporation on large areas if the lower air strata have uniform temperature and vapor content.

3.3.4 Radiation

Heat is the source of energy that causes evaporation. The primary source of heat is radiation from the sun. Thus the amount of evaporation is primarily dependent upon the amount of solar radiation received and retained by the water surface. As the amount of solar radiation is increased, the surface temperature of the water and vapor pressure differences between the water and the air are increased (Gray, 1970). This causes evaporation rates to rise.

3.3.5 Soluable Solids

When solutes are dissolved in water, the vapor pressure of the water is decreased (Chow, 1964). Lowering the vapor pressure of the water results in a reduced rate of evaporation. It was found that the rate of evaporation decreases as the specific gravity of salt water increases. Evaporation is about 2 to 3 percent less from sea water than from fresh water (Chow, 1964).

3.3.6 Depth of Water

The depth of water has an influence on the daily and seasonal variations in the rate of evaporation because of the ability of deep water bodies to store heat. In deep water a

portion of the radient energy received is stored in the water mass whereas it is not in shallow water bodies. This means that for an equal quantity of radiation received the amount of evaporation will be less for shallow water bodies than for deep.

3.3.7 Size of Evaporating Surface

The size of the evaporating surface affects the rate of evaporation. Over large surfaces of water there may be extreme variations of air and water temperature, as well as, wind and humidity, and consequently variations in the rate of evaporation. Also, air masses may be modified by humification as they travel over large surfaces of water (Hounam, 1973). This leads to a reduction in the vapor pressure deficit and, therefore, a reduction in evaporation.

3.4 EVAPORATION FROM SOIL

Atmospheric environmental conditions that affect evaporation from a free-water surface also affects evaporation from soils. The difference is with the nature of the surface from which evaporation takes place. The escaping molecules of water from soil must overcome the resistance due to the attraction of the soil particles for the water, (i.e. the field forces existing between the soil and water) (Chow, 1964). The end result is that the evaporation rate from soil is lower than that from open water.

The rate of evaporation from soil is dependant on the moisture content of the soil. However, unless the soil surface is kept moist, either by rainfall or irrigation or by rise of water from the water table, it will dry rapidly and evaporation will practically cease (Chow, 1964).

3.5 EVAPORATION FROM SNOW

Evaporation from snow may be treated fundamentally as evaporation from any other surface. However, certain properties peculiar to snow affect the net energy exchange and consequent evaporation rates.

Evaporation from snow involves a three-phase change of state from a solid to a liquid to a gas (Gray, 1970). This means that the potential energy for evaporation from snow is lower than from a free water surface due to the energy needed for melting. Frequently the mass of water melted is substantially greater than the amount evaporated (Gray, 1970). Also the snow surface has a higher reflectivity coefficient, thereby, lowering the potential energy available for evaporation.

3.6 THEORETICAL EXPLANATIONS OF EVAPORATION

There are three main theoretical approaches often used to explain evaporation. (1) The energy balance approach; (2) the mass transfer approach; and (3) the water budget approach.

3.6.1 Energy Balance Approach

The energy balance approach is based on the principle of the conservation of heat energy (Shiau, 1968). This method does not require any knowledge of the evaporation process but is calculated from consideration of the thermal budget of a body of water (Figure 3.1).

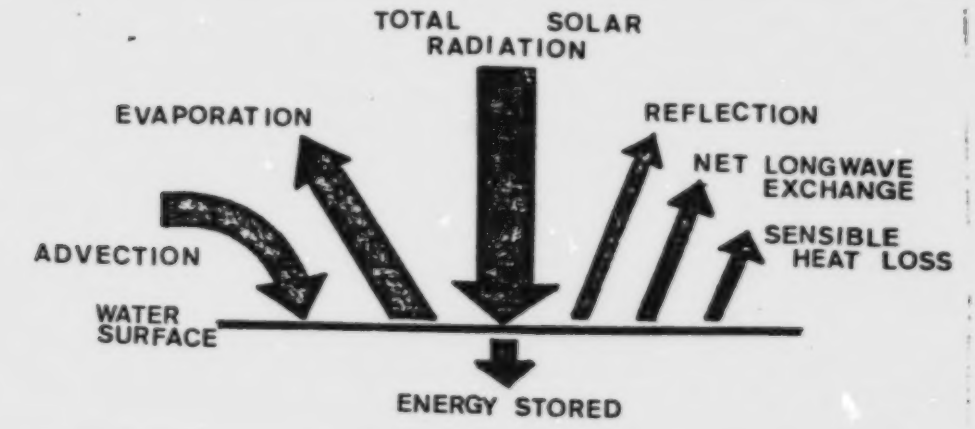

Figure 3.1 Energy budget over an open water surface

Estimates of the amount of energy needed to change water from a liquid phase to a vapor phase is used to determine the rate at which water is being removed.

The thermal budget may be expressed in equational form as:

$$R_o = R_t - R_r - R_l - R_s - R_E + R_v \qquad \text{(Eq. 3-1)}$$

where: R_o = increase in energy storage in the reservoir;
R_t = total solar radiation incident to the water surface;
R_r = reflected solar radiation;
R_l = net longwave radiation exchange between atmosphere and water;
R_s = sensible heat loss from the body of water;
R_E = energy utilized for evaporation;
R_v = net energy advected into the body of water;

This approach assumes that the amounts of energy used in the transformation of kinetic energy to heat, heating due to chemical or biological processes, and conduction of heat through the bottom of the reservoir are negligible (Gray, 1970).

By rearrangement of the energy budget equation, the energy utilized for evaporation can be quantified.

$$R_E = R_t + R_v - R_o - R_s - R_l - R_r \qquad \text{(Eq. 3-2)}$$

From this equation the amount of evaporation can be calculated as:

$$E = \frac{R_E = R_t + R_v - R_o - R_s - R_l - R_r}{L} \qquad \text{(Eq. 3-3)}$$

where L = latent heat of evaporation (cals/cm^3);
$L = 596 - 0.52T_w$; and
T_w = temperature of water surface (^oC)

Although this approach is sound in theory, its application is seriously restricted because of the difficulty in measuring the various components, and, for this reason, it is seldom used to estimate evaporation loss.

3.6.2 Mass Transfer Approach

The mass transfer approach to explaining evaporation is based on the determination of the mass of water vapor transfered from the water surface to the atmosphere. This movement of water is by eddy diffusion (Viessman, 1972). Eddy diffusion is evaluated based on the concepts of discontinuous and continuous mixing applied to mass transfer in the boundry layer (Shiau, 1968).

Many equations, both theoretical and empirical, have been developed. Most are similar in form to a relationship between evaporation and vapor pressure first recognized by Dalton in 1802 (Gray, 1970):

$$E = k(e_s - e_a) \qquad \text{(Eq. 3-4)}$$

where: E = evaporation in depth;
K = a coefficient dependent upon wind velocity and atmosphere pressure;
e_s = saturation vapor pressure at the water surface temperatures; and
e_a = vapor pressure of the air.

This equation illustrates two points: 1) the transfer of water vapor from the evaporating surface to the air occurs in the direction of the moisture gradient; and 2) the amount of evaporation is proportional to the product of wind speed and the humidity gradient.

The major advantage of this method is that all the factors involved can be directly measured or indirectly

evaluated from data. Numerous detailed mass transfer equations have been derived, some of which will be discussed later in this report.

3.6.3 Water Budget Approach

The water budget approach is based on the law of continuity. This law states that a change in storage of a water body is equal to the inflows minus the outflows (Kittredge, 1948). By determining these components the amount of evaporation can be determined. This relationship can be given in equational form as:

$$E = In + P - O - Og + S + Fi \qquad \text{(Eq. 3-5)}$$

where
- E = evaporation;
- In = inflow;
- P = precipitation;
- O = outflow;
- Og = seepage;
- S = change in storage; and
- Fi = stored ice.

This method has the advantage of being simple in theory but has a disadvantage of being difficult to prove because of the need for precise measurement of the terms of the equation. Any errors in measurement of the terms entering the equation are reflected in the computed amounts of evaporation. Evaporation as determined by this method is a residual, and may be subject to considerable error if it is small relative to the other items (Gray, 1970).

3.7 EVAPORATION REDUCTION

Reduction of evaporation is accomplished by controlling the rate at which water vapor escapes from water surfaces and may be of great economic importance. The methods of evaporation reduction may be classified into three groups: 1) those that reduce the surface area; 2) those that mechanically cover the area; and 3) those that cover the surface with a film.

The surface area of a reservoir can be reduced by constructing the reservoir with minimum ratio of area to storage (Chow, 1964). This can be accomplished by storing water below ground or storing water in one large reservoir instead of several small ones.

Using mechanical covers to reduce evaporation applies mostly to small reservoirs. Covers may include: roofs; floating rafts; and windbreaks (Gray, 1970). The mechanical cover reduces the amount of solar radiation reaching the surface of the reservoir and therefore decreases the evaporation rate.

The use of films to reduce evaporation appears to be the most economical and offers the greatest potential for reduction (Chow, 1964). Films of monomolecular thickness are spread on the surface of the reservoir to reduce evaporation.

Hexadiconal is a compound which forms a monomolecular film when poured on the water surface. The film formed by hexadiconal is only six ten-millionths of an inch thick

(Viessman, 1972). The film is penetrable by raindrops but offers a barrier to prevent water molecules from escaping.

The effectiveness of the monofilm depends on many factors, such as solubility, wind action, method of application and degradation by micro organisms (Chow, 1964). Each of these factors must be evaluated separately for each location in which this method might be used.

3.8 MEASUREMENT OF EVAPORATION

The difficulty in making direct measurements of evaporation from large water surfaces has led to the use of evaporation pans and tanks. These instruments attempt to integrate the effects of various climatic factors affecting evaporation and record a directly measureable loss that is proportional to that of an adjacent large water area (Meinzer, 1948). The advantage of using these instruments for estimating evaporation stems from the fact that they are inexpensive, simple to instrument and the annual ratio of lake to pan evaporation remains reasonably constant from year to year (Gray, 1970).

3.8.1. Exposure of Pans and Tanks

Evaporation from a pan and its variability over a period of time depends on the exposure of the instrument. The type of pan exposures that may be used include: the sunken pan; the floating pan; and the surface pan.

The sunken pan is placed in the ground with the rim of the pan at or near ground level. It has the advantage that certain boundary effects, such as direct solar radiation on the sides of the walls, are negligible (Gray, 1970). However, this advantage is sometimes outweighed by disadvantages. They tend to collect trash; are difficult to maintain and install; and leaks are hard to detect.

In the floating type of exposure the pan is floated on a lake or reservoir. This exposure gives the best estimates of evaporation because the water in the pan is under the same conditions as the water in the reservoir. The use of this type of exposure has not been widespread, however, due to problems of inacessability and water loss by wave action.

The surface pan exposure is the most common type of exposure used. In this exposure a pan is normally mounted on a wooden platform with an air space underneath. This eliminates difficulties caused by drifting soil and snow and at the same time permits air circulation under the pan (Gray, 1970). Since the rate of evaporation is strongly influenced by wind and water temperature, the pan is usually operated with an anemometer near the surface of the water and a maximum/minimum thermometer in the water (Lee, 1980).

3.8.2 Types of Evaporimeters

There are several types of evaporimeters in use throughout the world (Table 5.1). These instruments come in a variety of shapes, sizes and exposures which make it difficult to compare evaporation data from location to location. The four most commonly used evaporimeters include: 1) GGI-3000 pan; 2) 20m^2 tank; 3) Experimental Farms Buried Tank; and 4) U.S. Weather Bureau Class A Pan.

3.8.2.1 GGI - 3000 Pan

This instrument has a cylindrical tank with a conical base usually made of galvanized sheet iron (Figure 3.2). The diameter is 61.8 cm, (24.3 in) the depth is 60 cm (23.6 in) at the walls (Hounam, 1972). The tank is sunk in the ground or installed on a raft floating on a lake with the rim approximatley 7 cm (2.7 in) above the surface. Measurement is made using a fixed point gauge.

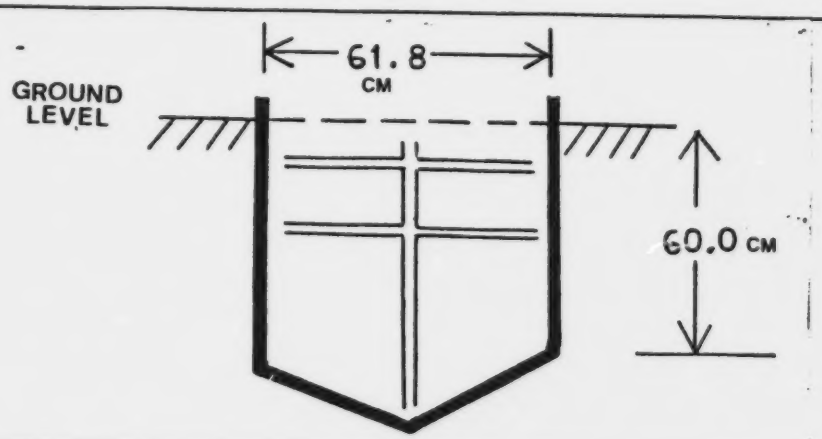

Figure 3.2 The GGI-3000 evaporation pan

TABLE 3.1 VARIOUS EVAPORIMETERS IN USE THROUGHOUT THE WORLD

Type	GGI-3000	20 m^2	U.S. Class A	Colarado Sunken Pan	MO Tank	Australian Sunken Tank	Anzac Tank	Experi-mental Farms Tank	U.S. Geol. Survey Pan	BPI Pan
Exposure	Sunken/ Floating	Sunken	Surface	Sunken	Sunken	Sunken	Sunken	Sunken	Floating	Sunken
Shape	Circular	Circular	Circular	Square	Square	Circular	Circular	Circular	Circular	Square
Side Length/ Diam (cm)	61.80	500.00	121.00	91.40	183.00	91.40	91.40	120.0	90.0	183.00
Side Length/ Diam (in)	24.30	196.80	47.60	36.00	72.00	36.00	36.00	47.2	35.4	72.00
Depth (cm)	68.50	200.00	25.40	45.70	61.00	91.40	91.40	60.0	45.0	61.00
Depth (in)	27.00	78.70	10.00	18.00	24.00	36.00	36.00	23.6	17.7	24.00
Volume (m^3)	2.01	38.80	0.29	0.38	2.02	0.59	0.59	0.67	0.282	2.02
Volume (ft^3)	7.25	1385.40	10.30	13.5	72.0	21.21	21.21	23.90	10.08	72.00
Material	Sheet Iron	Boiler Plate Sheets	Iron/ Fiber-glass	Iron	-	Iron/ Copper	Concrete	Iron	Iron	Iron
User	USSR	USSR	USA	USA	UK	Australia	Australia	Canada	USA	USA

3.8.2.2 20 m^2 Tank

This tank is cylindrical with a flat base and is normally made with boiler plate sheets of metal (Figure 3.3). The diameter is 5 m (196.8 in), surface area, 20 m^2, and the depth 2 m (78.7 in) (Hounam, 1972). The tank is installed in the ground with the rim approximately 7.5 cm (2.9 in) above ground. The water is normally maintained at ground level. Measurements are made using a gauge tank.

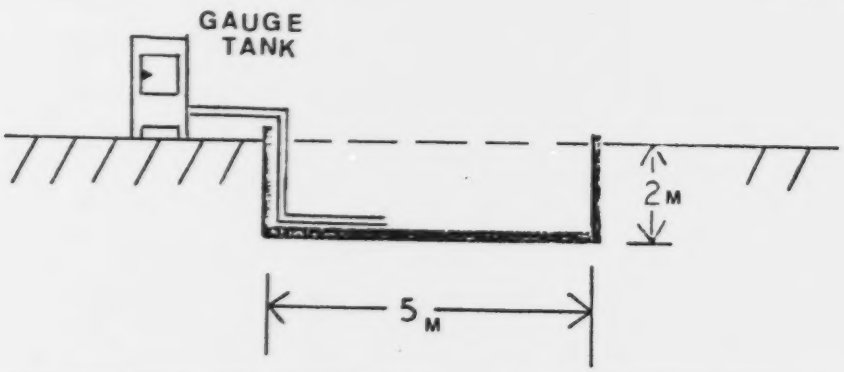

Figure 3.3 The 20 m^2 evaporation tank.

3.8.2.3 Experimental Farms Buried Tank

This type of tank has been used at the Canadian Department of Agriculture's experimental farms for more than sixty years. It is 4 ft (120 cm) in diameter 2 feet (60 cm) deep and installed with the rim 2 in (5 cm) above ground level (Pelton, 1964). (Figure 3.4). The water is maintained between 2 and 4 inches (5 and 10 cm) below the rim. Daily evaporation is determined using a micrometer gauge .

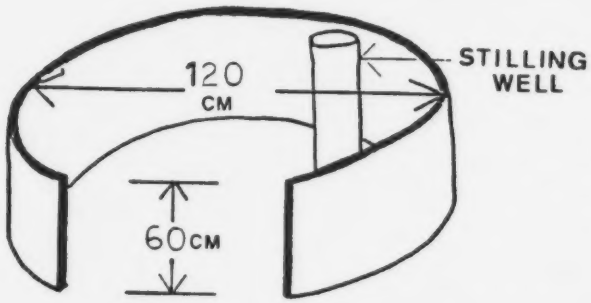

Figure 3.4 Experimental Farms Buried Tank.

3.8.2.4 U.S. Weather Bureau Class A Pan

This type of pan is the most widely used in North America. It is circular with a diameter of 47.5 inches (121 cm) and is 10 inches (25.4 cm) deep (Hounam, 1972) (Figure 3-5) It is mounted on an open wooden platform so that the base is about 2 to 4 inches (5-10cm) above the ground.

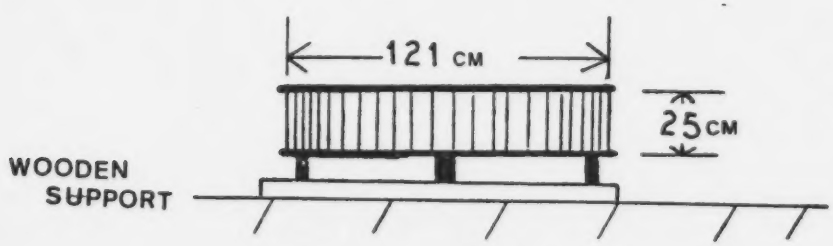

Figure 3.5 U.S. Weather Bureau Class A Evaporation Pan.

It is normally constructed of iron but other metals or fiberglass have been used. Measurements are made by hook or fixed point gauges.

3.8.3 Evaporation Pan Coefficients

The amount of evaporation that is measured from a pan must be corrected to give the amount of evaporation from a reservoir surface. The evaporation pan coefficient is the adjusting factor. The pan coefficient is given by a ratio of pan evaporation to lake evaporation. The value of the pan coefficient ranges between 0.60 and 0.80 for the U.S. Weather Bureau class A evaporation pan. The pan coefficient is multiplied by the amount of evaporation from the pan to obtain an estimate of lake evaporation.

Pan coefficients are not constant but vary from season to season, as well as, location to location. For this reason there is no simple universal index of conversion from pan evaporation to lake evaporation. True values of pan evaporation for a given period of time can only be obtained as a result of specific studies such as those involving the energy balance method or mass transfer method (Hounam, 1972).

3.8.3.1 Seasonal Variation of Coefficients

Large seasonal variations occur in monthly values of the pan coefficients. The chief factor causing the seasonal variation is the storage of heat in a large body of water

which results in evaporation not being in phase with solar radiation (Hounam, 1972). Evaporation from pans, however, are almost always in phase with radiation and hence the pan coefficient must be changed with season to obtain an accurate estimate of lake evaporation.

Another factor contributing to the seasonal variation is the differential advective heat transfer of pans or tanks compared with a lake (Hounam, 1972). The relative amounts of advected heat between a lake and pans are different seasonally because of the differences between pan air and water temperature and between lake air and water temperature.

3.8.3.2 Spatial Variation of Coefficients

Evaporation pan coefficients are not transferable from one location to another. Climatic factors (which have been discussed earlier), such as wind speed, temperature, and vapor pressure deficit, influence the rate of evaporation and hence the ratio of pan to lake evaporation. Kohler, working with the Class A evaporation pan, showed a geographic variation of the pan coefficient ranging from 0.60-0.80. The highest values for the pan coefficients are in coastal areas where higher humidity and lower radiation exist. These higher values presumably would be applicable to coastal areas of other continents (Hounam, 1972).

3.9 EMPIRICAL METHODS OF ESTIMATING EVAPORATION

Many empirical and semi-empirical equations have been developed to estimate evaporation from free water surfaces. Most of the formulas are based on the mass transfer theory of evaporation. Other formula rely on the conservation of energy theory or a combination of these two.

The greatest appeal of empirical formula is the fact that they permit estimates of evaporation to be made from standard meteorologic data (Gray, 1970). However, many require water surface temperature data which in practice is very difficult to obtain. Also, over-estimation or under-estimation of results may occur due to averaging of the meteorological data.

3.9.1 Mass Transfer Equations

Mass transfer equations are based on the mass transfer theory of evaporation and Dalton's aerodynamic law. Numerous equations have been derived using this approach. In most equations, evaporation is assumed proportional to a function of average wind speed and the difference between the vapor pressure of the air at the evaporating surface and the vapor pressure of the air at some level above the surface.

3.9.1.1 Meyer Equation

This mass transfer equation gives monthly estimates of evaporation. Meyer's work is based on fifty years of U.S.

Weater Bureau records. An empirical equation in the following form was developed.

$$E = k(1 + 0.1W)\ (e_s - e_a - 0.012) \qquad \text{(Eq. 3-6)}$$

where
- E = evaporation (inches/month);
- k = coefficient accounting for various factors (k=15 for large deep lakes and k=11 for small shallow lakes);
- W = mean montlhy wind velocity (mph a 25ft);
- e_s = saturated vapor pressure in inches of mercury corresponding to mean monthly air temperature for small water bodies or corresponding to water surface temperature for large bodies of water (Appendix I, Table A);
- e_a = actual vapor pressure in air based on monthly mean air temperature and relative humidity, (eq = 0.01 RH(e_s): RH in %).

The data required for application of the Meyer equation are: 1) air temperature; 2) water surface temperature; 3) wind speed; and 4) relative humidity.

The main limitation of this method is the lack of accurate water surface temperatures. Small bodies of shallow water are considered to have a water surface temperature regime similar to that of the air. An estimate of water surface temperature for large deep lakes may be obtained by using regression equations.

3.9.1.2 Rohwer Method

Rohwer's equation is also based on the mass transfer theory of evaporation and gives daily estimates. Rohwer derived the following formula as a result of extensive

investigational work at the Bureau of Agricultural Engineering of the U.S. Department of Agriculture:

$$E = 0.771(1.465 - 0.0186B)(0.44 + 0.118W)(e_s - e_a) \quad \text{(Eq. 3-7)}$$

where E = evaporation (inches/day);
B = mean barometric pressure (inches Hg at 32 F);
W = mean velocity of wind at the surface (mph);
e_s = mean vapor pressure at water surface temperature (inches Hg); and
e_a = actual vapor pressure of air (inches Hg).

The data required to use this formula include: 1) daily mean wind speed; 2) air temperature; 3) water surface temperature; and 4) relative humidity.

As for the Meyer equation, the major limitation of this method is the difficulty in obtaining the surface temperature of the water. However, Rohwer showed that the difference of computed evaporation from the observed evaporation during any month ranged from twelve to fifteen percent when the water surface temperature was equated to air temperature (Chow, 1964).

3.9.1.3 Thornthwaite - Holzman Method

Thornthwaite and Holzman developed a third equation to compute the amount of evaporation based on the mass transfer theory. The equation assumes an adiabatic atmospheric condition and logarithmic distribution of wind speed and mositure in the vertical (Chow, 1964). Hourly evaporation rates may be calculated using this method. The equation is:

$$E = \frac{133.3\ (e_1 - e_2)\ (w_2 - W_1)}{(T - 459.4)\ \ln\ (h_2/h_1)^2} \qquad \text{(Eq. 3-8)}$$

where: E = evaporation (inches/hour);
e_1 = vapor pressure at h_1 (inches Hg);
e_2 = vapor pressure at h_2(inches Hg);
w_1 = wind velocity at h_1 (mph);
w_2 = wind velocity at h_2 (mph);
h_1 = lower height above surface (feet);
h_2 = upper height above surface (feet); and
T = mean air temperature between h_1 and h_2 (^oF)

The data required to use this equation include: 1) mean air temperature; 2) wind velocity at two heights; and 3) relative humidity. This equation can only give precise results of hourly evaporation when the instrumentation for gathering the data meet exacting requirements (Chow, 1964).

3.9.2 Energy Balance Method

Energy balance equations used to estimate evaporation from free water surfaces are based on the conservation of energy theory. This method is normally applied to obtain estimates of annual evaporation from large water bodies (Chow, 1964). All energy balance equations account for incoming and outgoing energy, balanced by the amount of energy stored.

3.9.2.1 Anderson Equation

The Anderson equation assumes the principle of conservation of energy but neglects items of small magnitude. Estimates of annual evaporation obtained.

$$E = \frac{R_s - R_r - R_l - R_v - R_o}{\rho L \left[1 + \left(\frac{0.61P}{1000} \times \frac{T_w - T_a}{e_s - e_a}\right)\right]} \qquad \text{(EQ. 3-9)}$$

where: E = evaporation (cm);
R_s = solar radiation incident to water surface (cal/cm^2-day);
R_r = reflected solar radiation (cal/cm^2-day);
R_l = net long-wave radiation exchange (cal/cm^2-day);
R_v = net energy advected into body of water (cal/cm^2-day);
R_o = increase in energy storage (cal/cm^2-day);
ρ = density of water;
B = atmospheric pressure (mb);
L = latent heat of evaporation;
T_w = water surface temperature (^{o}C);
T_a = air temperature (^{o}C);
e_s = saturated vapor pressure (mb); and
e_a = vapor pressure of the air (mb).

The data needed to calculate evaporation by this method are extensive. All the radiation factors, including total and reflected radiation, net longwave exchange, sensible heat transfer, and increase in energy storage, must be evaluated, along with air temperature, water temperature and the vapor pressure deficit. For this reason, the use of the energy budget method is limited.

Accuracy of estimates obtained by using this method are highly dependant upon the reliability and accuracy of measured data. Under good conditions average errors of 10% for summer and 20% for winter can be expected (Viessman, 1972).

3.9.3 Penman Combination Method

The combination method for estimating evaporation from meteorologic data uses (as the title suggests) a combination of the mass transfer method and the energy balance method in one equation. The main purpose of this approach is to eliminate the need for the surface temperature of the water.

Penman with the use of two new parmeters (Ea and), developed the equation:

$$E = \frac{\Delta/\gamma\ H + E_a}{\Delta/\gamma + 1} \qquad \text{(Eq. 3-10)}$$

where: E = evaporation (mm/day);
E_a = expression of the drying power of the air (mm/day);
H = heat budget;
Δ = slope of the saturation vapor pressure curve at mean air temperature (Appendix I, Figure A); and
γ = constant = 0.27 for temperature in ^{O}F and vapor pressure in mm Hg.

E_a can be claculated as:

$$E_a\ (0.136 + 0.095\ W_2)\ (e_s - e_a) \qquad \text{(Eq. 3-11)}$$

where e_s = saturated vapor pressure (millibars) (Appendix I, Table A):
e_a = 0.01 RH(e_s) RH in %; and
W_2 = monthly wind speed (m/s @ 2m)

The heat budget (H) takes into account the incoming short-wave radiation from sun and sky, and the long-wave exchanges between earth and sky (Shiau, 1968). Mateer developed an equation to evaluate (H) based on Canadian

conditions. The equation may be expressed as:

$$H = (1-r)R_A(0.355+0.68\ n/N) - \sigma T_a^4(0.56-0.09\ e_a)^{1/2}(0.1+0.9\ n/N) \quad \text{(Eq. 3-12)}$$

where r = radiation reflection coefficient (0.05-0.15 for H20);
R_A = thearetical maximum solor radiation that reaches the site (Appendix I, Table B);
n/N = ratio of actual to possible hours of bright sunshine (Appendix I, Table C);
σT_a^4 = thearetical black body radiation at mean air temperature (Ta) (mm H20/day) (Appendix I, Table D); and
e_a = actual vapor pressure in air (inches Hg).

The data that needs to be measured in order to use this approach include: 1) duration of bright sunshine; 2) mean air temperature; 3) mean relative humidity; and 4) mean wind speed.

This method is sometimes limited in application because of the many meteorological observations required and because it is necessary to use charts and tables.

This method should not be used in arid regions because the presence of advected energy is not accounted for (Penman, 1956). The value of the heat budget dominated the expected value of evaporation, and the accuracy of the drying term Ea does not need to be as high as that of the heat budget (H).

CHAPTER 4
TRANSPIRATION

4.1 GENERAL

Transpiration is defined as the process whereby the water that has circulated through a plant structure is returned to the atmosphere principally in the form of water vapour (Brater, 1967). The transpiration process constitutes an important phase of the hydrologic cycle since it is the primary mechanism by which precipitation falling on land areas is returned to the atmosphere (Linsley, 1975). When studying the water balance of a drainage basin, evaporation and transpiration are normally treated as a single factor. However, a knowledge of the process of transpiration is required to assure that the techniques employed to measure evapotranspiration are consistent with physical reality.

4.2 THE TRANSPIRATION PROCESS

The water available for transpiration comes from a continuous column which extends from the roots in the soil through the stem and into the leaves. The movement of water in this continuous column is explained by the tension-cohesion theory. By this theory a column of water in a thin capillary tube is pulled up to considerable heights because of the tension transmitted down the tube and the cohesive

properties of the water molecules (Chow, 1964). As the plant transpires and the water content of their cells is lowered, they develop a reduced water potential (suction) which enables them to withdraw water from the soil.

The surface for transpiration is largely in the leaves. Water transferred through the plant is deposited in the intercellular spaces within the leaves and from there diffuses to the substomatal air spaces beneath each stoma (Rutter, 1972). The stoma is a tiny long opening through the epidermis of the leaf. Each stoma lies between two guard cells, which in turn are bordered by subsidiary cells (Figure 4.1). Opening and closing of the stoma result from unequal thickening of the guard cell walls which is caused by changes in the turgar of the guard cells.

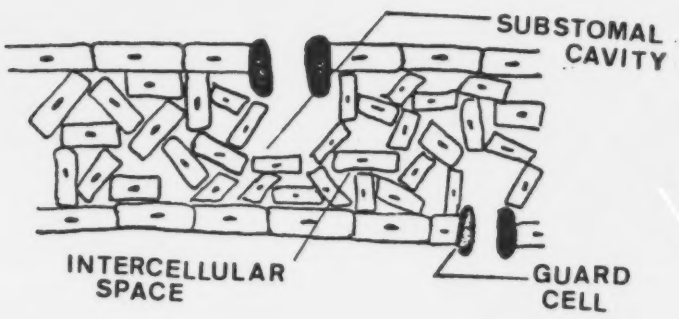

Figure 4.1 Cross-section of a leaf showing the guard cells, substomal cavity and subsidiary cells.

The stomata exist on either one or both sides of the leaf (but more frequently on the underside). When the stomata are fully opened the transpiration rate is determined by the same factors that control evaporation alone (Viessman, 1972). The stomata exert a slight regulatory influence only when they are almost completely closed.

4.3 FACTORS INFLUENCING TRANSPIRATION

The rate of water loss from a plant by transpiration is dependent on a combination of many factors. It varies greatly with the species of plant, the condition under which it grows, the time of day, month, or year, and from year to year.

The factors can be categorized as being physiologic, in which the plant itself influences the rate of transpiration or environmental in which the surrounding atmospheric conditions influence the rate.

4.3.1 Physiologic Factors

The physiologic factors that control transpiration include: plant type, density of stomata and leaf structure.

Plant type becomes an important factor in controlling transpiration when available soil moisture is limited. As the upper layers of the soil dry out, shallow rooted species can no longer obtain water and wilt, while deep rooted species continue to transpire until soil moisture at greater

depths is reduced to the wilting point (Linsley, 1975). Thus deep rooted vegetation transpires more water during sustained dry periods than shallow rooted species.

Plant type also influences the rate of transpiration when soil moisture is not limited. For example xerophytes, desert species, transpire relatively little water, whereas Phreatophytes, deep rooted species, transpire at rates largely independent of soil moisture (Linsley, 1975).

The number of stomata per unit area of a leaf is a second physiologic factor which affects transpiration. There is an increase in the rate of transpiration with an increase in the density of stomata. The density of stomata is characteristic of the plant species and ranges between 50,000 and 800,000 per square inch (8,000 - 128,000 /cm^2) (Chow, 1964). Density also varies in different leaves on the same plant (Meinzer, 1948).

Leaf structure has also been considered to have an effect upon transpiration. Transpiration is much less from leaves with dense palisade spongy tissue with small intercellular spaces and deep set stomata, as compared to leaves with sparse palisade tissue and large intercellular spaces (Meinzer, 1948). Another feature of leaf structure that affects transpiration is the incipient drying rate of cell walls. Incipient drying is the retreat of water films into the pores of the cell walls with an increase in the evaporating power of the air. If the water content of the

leaves is high, increased power of evaporation does not produce incipient drying, and transpiration increases. If, on the other hand, the water content of the leaves is low, incipient drying occurs and the rate of transpiration tends to decrease (Linsley, 1975).

4.3.2 Environmental Factors

The principal environmental factors that influence transpiration include sunlight, soil moisture and wind.

Sunlight has a major influence on the rate of transpiration. Transpiration follows a diurnal cycle which corresponds to that of light intensity (Viessman, 1972). This diurnal cycle is evident principally because transpiration varies directly with plant growth, which is almost wholely dependent upon sunlight (Brater, 1967). As a result, transpiration is virtually restricted to daylight hours. Increased sunlight also causes increased transpiration due to higher temperartures of the leaves. These higher temperatures result in greater evaporation from the substomatal cavities.

Transpiration is limited by the rate at which soil moisture becomes available. However, literature dealing with the effects of varying amounts of soil moisture upon transpiration and plant growth show conflicting evidence. Some investigators believe that transpiration and plant growth are affected in rough proportion to a decrease in soil moisture content until the wilting point is reached at which

time transpiration ceases (Chow, 1964). Others believe that transpiration and plant growth are unlimited by a reduction in soil moisture until the wilting point is reached (Linsley, 1975). It is believed that this controversey can be attributed to the various methods of obtaining data.

Wind movement is thought to increase transpiration. Wind carries away the moist air which forms under the leaf and replaces it with drier air, thus increasing the vapor pressure difference between the inside and outside of the leaf and increasing transpiration (Chow, 1964). A gentle wind is relatively much more effective in increasing transpiration than are winds of greater velocity. The activity of leaves increased with wind velocity up to 8 m.p.h. (13 km/hr) and thereafter the rate of transpiration becomes less (Chow, 1964). Gentle air currents stimulate transpiration but strong air currents may inhibit the process.

4.4 MEASUREMENT OF TRANSPIRATION

There are a variety of methods used to obtain estimates of transpiration. They either follow the loss in weight of a plant soil system with the surface of the soil covered, or they measure the water vapor evolved from the plant or some of its leaves, or they deal with watershed studies.

Most of these methods only give data for a few hours or days determined using detached parts of plants. For this

reason they are most often used by botanical researchers investigating transpiration as a physiologic process. Engineers and agriculturalists who require data for longer periods, ranging from a few weeks to the full growing season, and who are interested in transpiration of extensive stands rather than a part of a plant, seldom use these methods.

4.4.1 **Measurement of Water Vapor Transpired**

Direct measurement of the water vapor transpired from a leaf or plant usually requires enclosure in a chamber and this almost always alters the conditions in which transpiration occurs. Although this may limit the usefulness of the data, the technique is well suited to investigating the responses of transpiration to experimentally induced alterations in air temperature, humidity, or wind speed (Rutter, 1972). The three most popular methods of measuring the water vapor transpired are the hygroscopic absorption method, hygrometer method and indicators applied to the leaf method.

4.4.1.1 Hygroscopic Absorption

This method uses a closed container and moisture absorbing chemicals to get a measure of transpiration. The instrument consists of a glass cylinder, as a transpiration chamber for a growing twig, two U tubes containing phosphoric anhydride for absorption of moisture and a water aspirator of

known capacity all connected by rubber tubing (Meinzer, 1948). The amount of moisture absorbed by the U tubes is measured when air is drawn through the system by opening the aspirator valve. The leaf area in the cyclinder is then determined and the rate of transpiration calculated.

4.4.1.2 Hygrometer

This method consists of measuring changes in humidity in a closed container in which is placed a small growing plant or twig. The increase in humidity over time is related to the rate of transpiration. A problem arises with the use of this method, however, in that abnormally high humidities inside the chamber have been observed. For this reason the method is normally only used to give an index of transpiration.

4.4.1.3 Indicators Applied to the Leaf

This method utilizes strips of filter paper which change colour when moisture is applied. These filter papers are treated in a solution of salt then placed against a growing leaf. The time required for colour change, called leaf test time, is obtained and compared with the water test time for a standard evaporating surface, to obtain an index of transpiration power. This method is very sensitive to changes in transpiration rates and is normally used to compare rates between species.

4.4.2 Measurement of change in weight

Methods that measure a change in weight of a plant soil system are indirect methods of determining transpiration. The change in weight is due to a loss of water from the plant and therefore is related to transpiration. These methods are often used to determine the effects of environmental conditions upon growing plants or they are used to compare transpiration rates between species. The three most commonly used methods are the potometer, the phytometer, and the transpiration ratio.

4.4.2.1 Potometer

The potometer consits of a vessel containing water which can hold the cut end of a leaf, twig, or branch. After proper sealing, measurements are made of the amount of water removed from the vessel by periodic weighing of container and speciman. This method is not adapted to obtaining reliable quantitative results applicable to plants normally rooted in the soil. It is most often used by botanists for theoretical laboratory studies of transpiration.

4.4.2.2 Phytometer

This instrument consists of a watertight tank filled with soil in which one or more plants are rooted (Chow, 1964). A cover is sealed over the instrument to prevent any water from escaping except through transpiration (Linsley, 1975). The

transpirational losses for any period are equal to the original weight plus the weight of the water added minus the final weight. Successive weights are obtained by automatic balances in a laboratory. It is the most satisfactory and convenient method for studying transpirational losses although its application is limited to plants with small root systems (Linsley, 1975).

4.4.2.3 Transpiration Ratio

This method is a modification of the phytometer method and is used to determine the total amount of water lost in relation to dry matter produced (Brater, 1967). The ratio of the weight of water consumed, and therefore transpired, and the weight of dry matter produced is called the transpiration ratio (Linsley, 1975). The transpiration ratio serves primarily as a measure of relative efficiency of plants in the use of water. In some places, where data on crop production are available, it can be applied to determine transpiration loss from large areas of cropped land (Viessman, 1972).

The method consists of growing plants to maturity in sealed tanks containing about 136 kg of soil for a specified time interval. At the end of the test, the above ground plant is oven dried and the weight of dry matter produced is calculated. The amount of water consumed by the plant is computed from the sum of the quantities of water added during

the season corrected for the difference in weight of the container with contents at the beginning and end of the test (Chow, 1964).

4.4.3 Watershed Studies

Another method of determining transpiration is by means of watershed studies. In this method the vegetative cover of the watershed is removed and the apparant increase in runoff is computed (Chow, 1964). The method requires a period of intensive study of rainfall and runoff before and after the vegetation is removed, or requires a correlation of runoff between the basin being studied and an adjacent control basin (Baier, 1967). The increased runoff from the clear-cut basin is used as an estimate of transpiration.

It may not be possible to determine exactly a corresponding change in transpiration because the rate of soil evaporation may have also undergone a change. Therefore the watershed studies method only gives a general estimate of transpiration (Baier, 1967).

CHAPTER 5
EVAPOTRANSPIRATION

5.1 GENERAL

The evapotranspiration component of the hydrologic cycle includes water lost from land areas by evaporation and transpiration. It is defined as evaporation from all water, soil, snow, ice, vegetation and other surfaces, plus transpiration (Chow, 1964). Approximately 70 per cent of the precipitation falling on land areas is returned to the atmosphere by evapotranspiration from soil and vegetation (Baier, 1967). Fluctuations in evapotranspiration rates are of interest to engineers studying watershed balances and agriculturists who are intereted in crop water requirements.

Although not strictly correct, many authors use evapotranspiration as a measure of consumptive use of water by plants (Chow, 1964; Gray, 1970; Linsley, 1975). Consumptive use equals evapotranspiration plus the water used directly to build plant tissue. The amount of water used in building plant tissue, is relatively small in comparison to evapotranspiration and, therefore, equating evapotranspiration and consumptive use is acceptable in most cases. Since the value of consumptive use falls well within the error of measurement of evapotranspiration, the two terms are treated as synonyms and used interchangably (Linsley, 1975).

5.2 POTENTIAL VS ACTUAL EVAPOTRANSPIRATION

Before any estimates of evapotranspiration can be made, a distinction between potential evapotranspiration (PE) and actual evapotranspiration (AE) must be made. Potential evapotranspiration (PE) is defined as the amount of water lost from an area by evaporation and transpiration if the soil moisture content is non limiting (i.e. moisture content is at field capacity) (Gray, 1970). The concept of potential evapotranspiration eliminates the need to know complex soil moisture effects and permits estimates of evapotranspiration based on meteorological data only. The rate of potential evapotranspiration is determined solely by atmospheric conditions and cannot exceed the rate of evaporation from an open water surface exposed to the same meteorological conditions (Baier, 1967).

Actual evapotranspiration (AE) is defined as the actual loss of water from an area due to evaporation and transpiration (Gray, 1970). The rate of actual evapotranspiration is dependant on soil moisture availability, plant physiological characteristics, and meteorological factors (Baier, 1967). The actual evapotranspiration rate can never be greater than the potential evapotranspiration rate, but is often lower because the soil moisture content is less than field capacity.

5.2.1 Relationship Between AE, PE and Soil Moisture

The difference between the actual and potential evapotranspiration rates under similar meteorological conditions is a result of the moisture content of the soil. As the water in the soil decreases from field capacity to the permanent wilting point, there is a decrease in the rate of actual evapotranspiration. This decrease ,however, is not fully understood and has long been a subject for debate (Baier, 1967). Three different theories have been developed to explain the decrease, although none are applicable under all soil and climatic conditions.

The first theory states that the actual evapotranspiration rate is independant of soil moisture until the permanent wilting point is reached (Veihmeyer, 1956; Van Bavel, 1960: Lowry, 1959). This means that the AE rate is equal to the PE rate over the range of soil moisture conditions to the permanent wilting point, at which time the AE rate drops off (Figure 5.1).

This explanation is acceptable for many sandy soils permeated with roots, and also in irrigation studies where water is added to maintain evapotranspiration at a maximum rate (Baier, 1967).

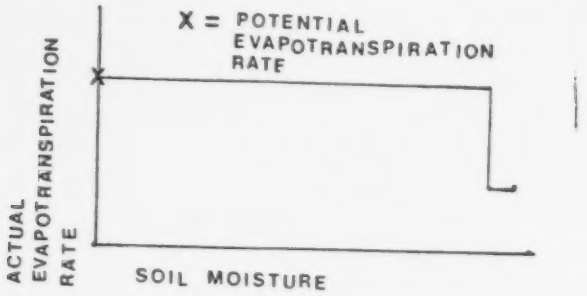

Figure 5.1. The actual evapotranspiration rate remains constant over the range of soil moisture conditions.

The second explanation states that there is no substantial reduction in actual evapotranspiration with soil moisture except in very dry soils (Figure 5.2).

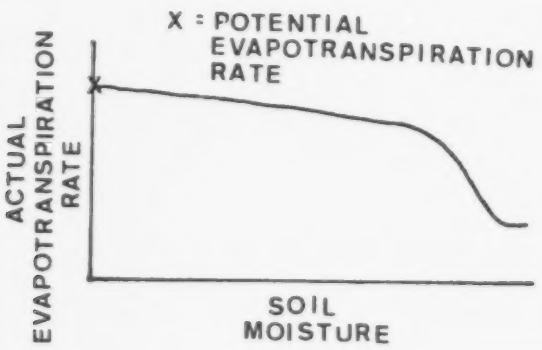

Figure 5.2. The actual evapotranspiration rate is maintained above 90 percent of the potential evapotranspiration rate until about 2/3 of the field capacity at which time the actual evapotranspiration rate declines.

It has been shown that the AE rate was maintained at approximatley 90 per cent of the PE rate until about two-thirds of available moisture has been exhausted (Pierce, 1958; Gardner, 1960; Shaw, 1968).

This concept applies to irrigated soils and to non-irrigated cropped soils in which a large part of available water can readily be withdrawn by plants (Baier, 1967).

The third hypothesis is that a linear relationship exists between actual evapotranspiration and soil moisture, (Figure 5.3). Thornthwaite and Mather (1955), through their work with sandy loam soils under high evaporative conditions, developed a linear relationship between AE and soil moisture. Other authors (such as Denmead and Shaw, 1962) have verified Thornthwaite's findings (Baier, 1967).

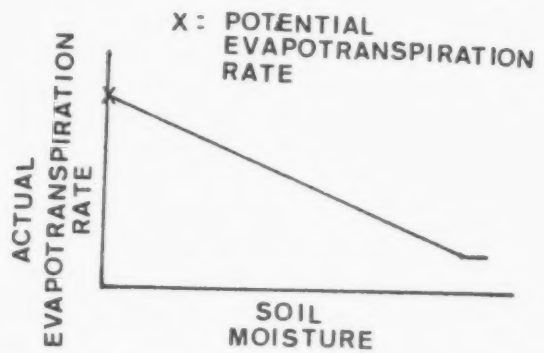

Figure 5.3. The actual evapotranspiration rate decreases linearly with soil moisture until the permanent wilting point is reached.

The preceeding discussion is given to show that the relationship between actual evapotranspiration (AE) and potential evapotranspiration (PE) is highly complex and difficult to establish for any area. The true relationship between AE, PE and soil moisture is most probably a combination of the three relationships given above.

5.3 FACTORS AFFECTING EVAPOTRANSPIRATION

The rate of evapotranspiration is influenced by a combination of the factors which affect both evaporation and transpiration. The factors which affect evaporation and trasnspiration have been discussed in detail in previous sections and the reader is, directed there for detailed information.

The meterological factors which affect evapotranspiration include: solar radiation, wind, relative humidity, and atmospheric pressure. These factors work in combination with vegetational and soil factors, such as plant type, density, stage of growth, and capillary and hydraulic conductivity of the soil, to determine the rate of evapotranspiration.

5.4 ESTIMATION OF EVAPOTRANSPIRATION

A variety of methods are available for estimating both actual evapotranspiration and potential evapotranspiration. No single method, however, can be used for all purposes. The

method that is chosen is dependant on the type of data that is required. For example, in some studies only annual evapotranspiration estimates are required, while in other cases, weekly water use by crops needs to be known.

The methods used to estimate evapotranspiration can be grouped into three categories: 1) obtaining direct measurements from which an estimate of actual evapotranspiration can be made; 2) estimating potential evapotranspiration from evaporation pans and atmometers; and 3) estimating evapotranspiration using empirical relationship and meteorological data. These methods are discussed in the following sections.

5.5 DIRECT MEASUREMENT OF EVAPOTRANSPIRATION

Direct measurement methods for estimating evapotranspiration involve the use of instruments which can measure inflows and outflows of water from a land area. These techniques give estimates of actual evapotranspiration, or potential evapotranspiration if the area under observation has sufficient water at all times. The three most widely used methods of direct measurement include: field plot methods; lysimetric methods; and watershed balance methods.

5.5.1 Field Plot Method

Determination of evapotranspiration by the field plot method simply involves measurement of inflow, outflow and

storage of water at a site. The amount of water added to a plot by precipitation and irrigation, as well as the water lost to surface runoff is determined. Evapotranspiration is then calculated as the difference between input and surface runoff and adjusted for changes in soil moisture storage (Gray, 1970). Changes in soil moisture storage are determined by direct measurement. Soil samples are taken in the plot before and after each irrigation and their moisture contents are determined by standard laboratory practices (Chow, 1964).

The primary advantage of this method is that reliable seasonal estimates of actual evapotranspiration can be made under field conditions. However, this method should only be used under ideal conditions. It should not be used on plots with deep percolate soils or on plots in which the water table is near the surface of the ground. Precise measurement of percolation is not possible using this method, and consequently the error it causes in the etimate of evapotranspiration tends to be significant (Linsley, 1975). The water table must be deep enough so that plants do not derive any of their water supply from this source (Gray, 1970).

5.5.2 Lysimetric Methods

Many observations of evapotranspiration are made in soil containers variously known as tanks, evapotranspirometers and

lysimeters. The instrument itself consists of a container with vertical sidewalls, open top, and a bottom that provides for percolation of water which is collected and measured (Figure 5.4).

The tanks are generally about 3 ft (1 m) in diameter and 6 ft (2 m) deep, although larger sizes up to 10 ft (3 m) in diameter and 10 ft (3 m) deep have been used (Chow, 1964). The lysimeter is filled with soil in which natural vegetation is grown. Evapotranspiration is determined from measurements of water applied, water lost through percolation and changes in soil moisture in the tank (Linsley, 1975). Changes in the soil moisture level may be obtained by direct moisture sampling, as discussed for the field plot method, or by using weighing and hydraulic lysimeters.

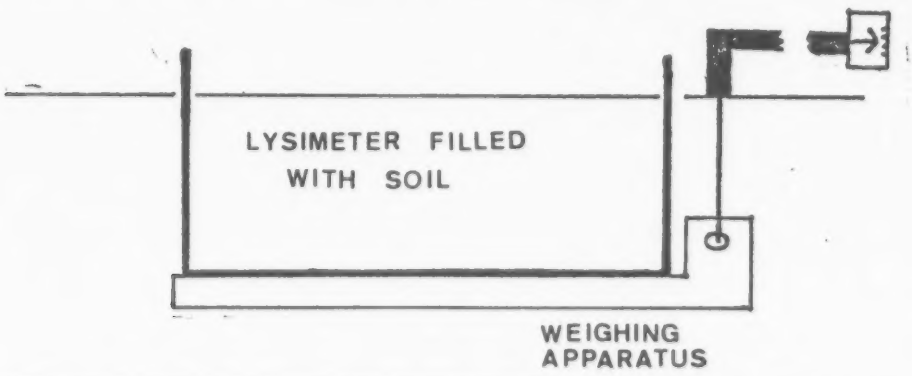

Figure 5.4. A simple lysimeter setup with a weighing apparatus to obtain a measure of soil moisture.

Under optimal conditions this method will provide reasonably reliable information on evaporation for short

periods of time such as an hour or day (Gray, 1970). However, lysimeters are often only used to provide an index to evapotranspiration because the overall condition of the tank may not closely simulate actual field conditions.

5.5.3 Water Balance Method

This method involves the application of the water balance principle to get estimates of annual evapotranspiration from large watersheds up to 200,000 hectares in area (Chow, 1964). Measurements of precipitation, streamflow, and surface runoff are made on a daily basis. Annual evapotranspiration is calculated as the difference between long term averages of precipitation and net stream flow (Baier, 1967).

The water balance method provides a convienient basic approach to analysing the overall water resources of a region. This method, however, is not suitable for calculating evapotranspiration from areas smaller than a watershed or for periods shorter than a year, when changes in moisture storage are important and cannot be evaluated.

5.6 ESTIMATING PE FROM ATMOMTERS AND PANS

An estimate of potential evapotranspiration from a land area may be obtained by evaluating the amount of water evaporated from an atmometer or evaporation pan. Evaporation from these instruments is used as an index to potential evapotranspiration. Evaporation pans are discussed in Chapter 3 and

the reader is referred there for a review of the application.

An atmometer is any instrument used to get a measure of the evaporative power of the air. They have a special surface that is kept moist and from which water loss is recorded. Atmometers are mainly used by agricultural workers who frequently require short term estimates of PE to determine irrigation requiremnts (Baier, 1967). They are rarely used to estimate evaporation from open water surfaces. The three most commonly used atmometers are the Bellani plate, the black porous disk, and the piche.

The Bellani plate atmometer consists of a thin porous black ceramic disk which is 3 inches (7.5 cm) in diameter and 1/8 inch (1/3 cm) thick which is fused to the top of an inverted glazed ceramic cup and suspended horizontally at a height of 4 feet (1.23 m) above ground level (Pelton, 1964). Water is conducted to the cup from a 250 cc burette which serves as a reservoir and measuring device (Figure 5.5). Maximum daily water loss in cubic centimetres is measured.

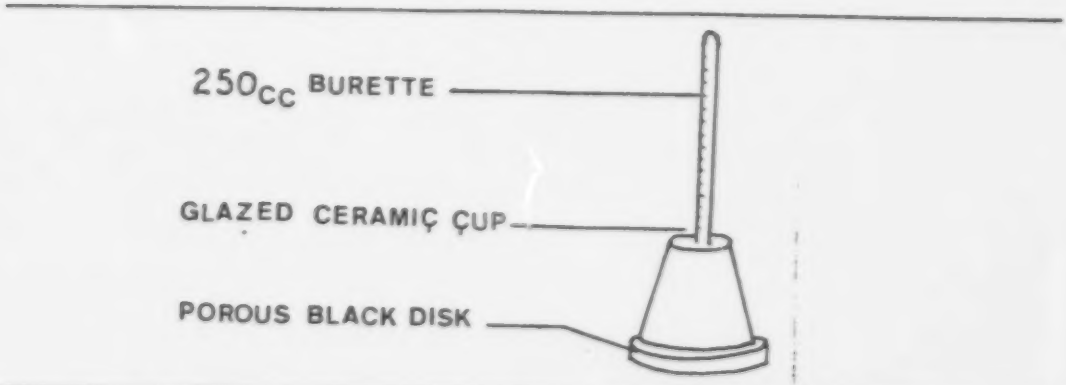

Figure 5.5. The Bellani plate atmometer.

The black porous disk atmometer consists of an aluminum disk which is 2 7/8 inches (7.19 cm) in diameter, 1/8 inch (1/3 cm) thick and supported in a horizontal position 4 feet (1.23 m) above ground level (Pelton, 1964). A reservoir of water is supported above the evaporating surface and water is supplied to the disk through a plastic tube (Figure 5.6). Evaporation is measured daily in a manner similar to that used with the Bellani plate atmometer.

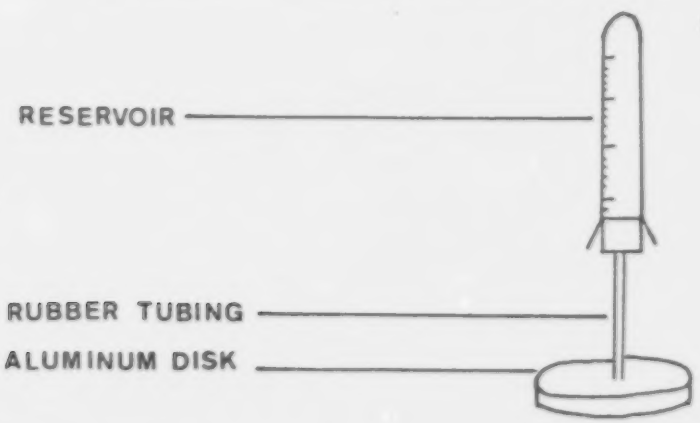

Figure 5.6. The black-porous disk atmometer

The piche atmometer consists of a graduated tube about 9 inches (22 cm) in length and 0.4 inches (1.0 cm) in diameter with one end closed and the other flat end covered by a circular piece of filter paper pressed against it by a disk (Chow, 1964)(Figure 5.7). The instrument is inverted and daily volume of water lost is measured.

An estimate of the potential evapotranspiration from a land area is calculated by applying an empirical coefficient to the reading taken from the atmometer or evaporation pan. These empirical coefficients are not constant, but vary from location to location and from season to season, as well as, from instrument to instrument. They must be derived from experimental research plots before this method can be used. The empirical coefficients vary over the entire growing season and take into account the degree of crop cover, crop roughness and color of vegetation (Baier, 1967).

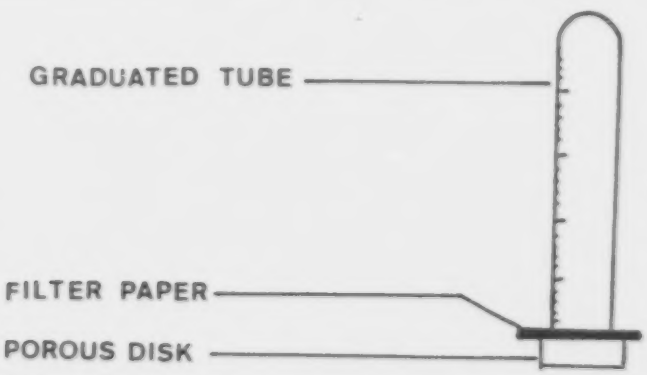

Figure 5.7. The piche atmometer.

Estimating evapotranspiration by this technique has the advantage of being inexpensive, easily maintained and easily standardized, however, there are also disadvantages. Only estimates of potential evapotranspiration can be obtained because the instruments cannot evaluate the effects of

varying soil moisture contents. Also errors in estimation often occur due to inadequate instrument exposure and incorrect empirical coefficients. However, this method, with correctly derived reduction constants, gives estimates of evapotranspiration which are sufficiently accurate in irrigation control (Stanhill, 1965).

5.7 EMPIRICAL EQUATIONS

Several empirical equations have been developed for estimating evapotranspiration from climatic data. The equations use a variety of climatic parameters ranging from air temperature to solar radiation to vapor pressure deficit to wind. Most of the equations can only give estimates of potential evapotranspiration because the equations cannot account for soil moisture factors. The principal equations most often used include: Thornthwaite equation; Blaney-Criddle equation; Papadakis equation; Penman equation; Lowry Johnson equation; and the Turc equation.

5.7.1 Thornthwaite Equation

The Thornthwaite equation is based on an exponential relationship between mean monthly air temperature and mean monthly evapotranspiration. Tne only factor that needs to be measured to apply this method is temperature. Temperature serves as an index to potential evapotranspiration because there is a fixed relation between the net radiation used for

heating and that used for evaporation (Baier, 1967).

Thornthwaite's relationship is based largely on experience in the central United States and is expressed as follows:

$$PE = 1.62\ a(10T/H_I) \qquad \text{(Eq. 5-1)}$$

where:
- PE = mean monthly potential evapotranspiration (cm);
- T = mean monthly temperature (C);
- H_I = annual heat index; and
- a = coefficient

The yearly heat index can be calculated from mean monthly air temperature by summing (H_I) for each month as:

$$H_I = (T/5)^{1.54} \qquad \text{(Eq. 5-2)}$$

The coefficient 'a' has a polynomial relationship with the annual heat budget and may be approximated by (Gray, 1970):

$$a = 6.75 \times 10^{-7} H_I^3 - 7.71 \times 10^{-5} H_I^2 + 1.79 \times 10^{-2} H_I + 4.9 \times 10^{-1} \qquad \text{(Eq. 5-3)}$$

Thornthwaite while developing this reltionship made two major assumptions: 1) a standard month of 30 days, each having 12 hours of possible sunshine; and 2) the range of temperature varies within 0 C and 40 C (Shaw, 1968). Since the number of days in a month varies from 28 to 31 and since the possible hours of sunshine vary with month and latitude,

the computed potential evapotranspiration value must be adjusted by a factor (Gray, 1970). The adjusting factors are shown in Appendix I, Table C. The major advantage of this method is that it is possible to arrive at a value of PE that is approximately correct using only temperature.

5.7.2 Blaney-Criddle Equation

The Blaney-Criddle method of determining evapotranspiration assumes that the heat budget is shared in fixed proportions between heating the air and evaporation (Gray, 1970). As a result of this assumption, mean monthly potential evapotranspiration is related to mean monthly air temperature and the monthly percent of annual daytime hours. The equation is:

$$PE = k \frac{TM_p}{100} \qquad \text{(Eq. 5-4)}$$

where:
PE = mean monthly potential evapotranspiration;
k = coefficient;
T = mean monthly air temperature; and
Mp = monthly percent of annual daytime hours.

The coefficient k varies with the type of vegetation, month, and locality (Baier, 1967). Therefore, to be able to use this method and obtain accurate results, k must be known for the time and place where the estimate of evapotranspiration is to be made (Chow, 1964).

To determine a seasonal value of potential evapotranspiration, the monthly values are summed over the growing season.

The Blaney-Criddle method should not be used for estimates of potential evapotranspiration for shorter periods than a month because no allowance is made for variations in wind and relative humidity.

5.7.3 Papadakis Equation

The Papadakis formula uses a relationship between vapor pressure and average daily maximum and minimum temperature to get an estimate of monthly potential evapotranspiration.

This formula is probably the simpliest for estimating potential evapotranspiration, and is:

$$PE = 0.5625 \, (e_{ma} - e_{mi-2}) \qquad \text{(Eq. 5-5)}$$

where:
- PE = monthly potential evapotranspiration in cm;
- e_{ma} = saturation vapor pressure (mb) corresponding to average maximum daily temperature (C); and
- e_{mi-2} = saturation vapor pressure (mb) corresponding to average daily minimum temperature (C) minus 2 C. (2 C is the normal difference between average daily minimum temperature and dew point temperature (Gray, 1970).

This formula integrates the effects of radiation advection and air dryness on potential evapotranspiration, when an average over 10 or more days is considered (Baier, 1967). The constant in the equation (0.5625) is used for all climates and for monthly potential evapotranspiration values throughout the year (Baier, 1967). This suggests that this method can be used in various locations with confidence.

5.7.4 Penman Equation

The Penman equation uses a combination of the energy balance theory and mass transfer theory of evaporation to give a daily estimate of potential evapotranspiration. The equation is virtually the same as the Penman equation for calculating evaporation from a open water surface, except various coefficients have been changed to meet forested conditions. The equation is:

$$PE = \frac{(\Delta/\gamma\ H + E_{at})}{(\Delta/\gamma + 1)} \qquad \text{(Eq. 5-6)}$$

where:
- PE = daily potential evapotranspiration;
- Δ = slope of saturation vapor pressure curve at mean air temperature (Appendix I, figure A);
- γ = constant (o,27 for temperature in C and pressure in mm Hg);
- H = heat budget; and
- E_{at} = evaporation rate from open water.

The heat budget (H) for the forested region at the North Nashwaak stream and other forest areas in New Brunswick can be evaluated by: (Shiau, 1968).

$$H = (1-r)R_a(0.36+0.68\ n/N) - \sigma T_a^4\ [0.56-0.092\ (0.01(RH)e_s)^{1/2}\ (0.1+0.9\ n/N] \qquad \text{(Eq. 5-7)}$$

where:
- r = radiation reflection coefficient (albedo);
- R_a = maximum solar radiation reaching a site if there were no atmosphere (Appendix I, Table b);
- n/N = ratio of actual to possible hours of bright sun (n is measured N see Appendix I, table c).
- T_a^4 = theoretical black body radiation at mean air temperature Ta (Appendix I, table d).

RH = mean relative humidity (mm Hg); and
e_s = saturation vapor pressure corresponding to mean monthly air temperature (Appendix I, table a).

The evaporation term (Ea) can be evaluated for forest areas in New Brunswick as:

$$Eat = 0.27(1 + 0.009W_{6.6})\ (e_s - e_a) \qquad \text{(Eq. 5-8)}$$

where: $W_{6.6}$ = monthly mean wind speed in mph at 6.6 feet above ground level;
e_s = saturated vapor pressure corresponding to mean monthly air temperature (Appendix I, table a);
e_a = $0.01\ (RH)e_s$; and
RH = mean relative humidity (Shiau, 1968).

In order to use the Penman method, duration of bright sunshine, mean air temperature, mean relative humidity, and mean wind speed must all be measured. The Penman method provides the best estimates of potential evapotranspiration. However, as a result of the many variables which need to be measured and the problem of defining coefficients for different locations, this method is seldom used (Gray, 1970).

5.7.5 Lowry-Johnson Equation

Lowry and Johnson developed a procedure for estimating annual potential evapotranspiration from agricultural lands. This method assumes a linear relationship between effective heat (accumulation in degree-days at maximum daily growing season temperature above 32 F) and potential evapotranspiration (Gray, 1970).

The relationship can be expressed as:

$$PE = a + kI_f \qquad \text{(Eq. 5-9)}$$

where: PE = potential evapotranspiration (acre-feet/acre);
a & k = coefficients; and
I_f = effective heat (thousands of degree-days).

The coefficients 'a' and 'k' vary from location to location, as well as, season to season and must be recalculated for the time and place where the method is to be used (Gray, 1970). 'a' and 'k' are evaluated using a simple linear regression of potential evapotranspiration on effective heat. The values of 'a' and 'k' most often used by the U.S. Bureau of Land Reclamation are 0.8 and 0.156 respectively (Gray, 1970).

5.7.6 Turc Equation

In Europe, Turc developed equations based on a statistical study of data collected from 254 watersheds located in all parts of the world. He related evapotranspiration to precipitation and temperature over watersheds (Gray, 1970).

The Turc equation can give annual estimates of potential evapotranspiration, when soil moisture is not limited, or annual estimates of actual evapotranspiration. Turc's equation relates soil moisture to precipitation and therefore actual evapotranspiration can be found.

The relationship is expressed by:

$$AE = \frac{P_a}{[0.90 + (\frac{P_a}{IT})^2]^{1/2}} \qquad \text{(Eq. 5-10)}$$

where: AE = annual actual evapotranspiration (mm);
P_a = annual precipitation (mm);
IT = 300 + 25T + 0.0573; and
T = mean air temperature (C).

The only data that needs to be collected are precipitation and temperature. The main advantage of this method is that an estimate of actual evapotranspiration can be made using only precipitation and temperature data. This method, however, is only reccommended when a general estimate of actual evapotranspiration is required because the equation does not take into account any local environmental effects.

CHAPTER 6
INFILTRATION

6.1 GENERAL

Infiltration is the process in the hydrologic cycle where precipitation that reaches the soil surface is passed into the soil. The maximum rate at which water can enter the soil at a particular point under a given set of conditions is called infiltration capacity (Linsley, 1975). Percolation is the movement of water within the soil and is closely related to infiltration since infiltration will not continue until percolation removes water from the surface of the soil. Generally, infiltration has a high initial rate that diminishes during continued rainfall to a nearly constant lower rate.

Infiltration affects many aspects of hydrology. Infiltration influences runoff; relates to transpiration of plants; and determines the moisture in the soil (Chow, 1964). The phenomenon of infiltration deserves special attention inasmuch as a thorough understanding of the process will enable the hydrologist to more effectively estimate amounts of runoff and thus to more confidently and competantly apply the results to design problems (Gray, 1970).

6.2 PROCESS OF INFILTRATION

Water reaching the soil surface enters the soil, (due to gravitational forces) through cracks and large pores which permeate the soil. Capillary forces continuously divert the water into capillary-pore spaces, so that the amount of water passing to lower horizons is steadily diminished. This leads to increasing resistance to gravity flow in the surface layer and a decreasing rate of infiltration as a storm progresses (Linsley et. al., 1975).

The rate of water entry into the soil is related to both the total porosity and size of the pores. Any factor which reduces either the number or size of pores reduces the rate at which the soil will absorb and transmit water (Gray, 1970).

Water cannot continue to enter the soil more rapidly than it is transmitted downward and, therefore, infiltration is highly dependant on the percolation rate. Percolation rates may vary for successive horizons of the soil profile. Any soil horizon may be naturally impermeable to water due to its texture and structure (Gray, 1970).

6.3 FACTORS AFFECTING INFILTRATION CAPACITY

The infiltration capacity of most soils is characterized by extreme variability. The actual value of infiltration capacity at any time or any location is a result of a combination of many factors. These factors include: 1) soil type;

2) soil moisture; 3) mechanical compaction; 4) characteristics of the fluid; 5) vegetation; and 6) season.

6.3.1 Soil Type

The type of soil determines the proportion of different sizes of pores, and their relative stability during storms and irrigation and, therefore, is a major factor affecting infiltration capacity (Chow, 1964). The infiltration capacity decreases as the texture of the soil becomes finer (i.e. as the soil changes from sand to silt to clay).

The soil type determines the amount of disintegration and swelling of soil particles. Soils with appreciable amounts of silt or clay are subject to disintegration of the aggregates into small particles which may clog pores and greatly reduce infiltration (Chow, 1964). The degree of swelling that occurs in soil is also affected by the kind of clay minerals present (Chow, 1964). With greater swelling there is a reduced infiltration capacity due to a reduction in pore size.

6.3.2 Soil Moisture

The amount of moisture in the soil affects the infiltration capacity by varying the capillary forces and storage capacity of the soil. If the soil is quite dry at the beginning of a storm, the wetting of the surface layer creates a strong capillary potential just under the surface which

supplements the gravitational force in causing infiltration (Wisler, 1967). When saturated, a soil has a relatively low infiltration capacity due to decreased storage capacity in the soil and decreased capillary potential beneath the surface (Chow, 1964).

6.3.3 Mechanical Compaction

Mechanical compaction decreases the infiltration capacity by creating an impervious layer at the surface of the soil.

Mechanical compaction caused by raindrops greatly reduces the infiltration capacity in soils of fine texture. The surface of exposed clay soils can be worked into an impermeable condition, due to rain, whereas the infiltration capacity of a clean sandy soil is affected very little by rain compaction (Wisler, 1967).

When heavy pedestrian traffic or animal passage occurs on a soil, the surface may become impermeable and the infiltration capacity lowered. Examples of types of areas which have low infiltration capacities as a result of mechanical compaction by people or animals are overgrazed pastures, playgrounds, and dirt roads (Wisler, 1967).

6.3.4 Characteristics of the Fluid

Another group of factors that affect infiltration are those that modify the physical characteristics of the water.

Impurities in the water have the effect of decreasing infiltration capacity. Many experiments have shown runoff to be less (and therefore infiltration to be greater) in areas where the water is relatively low in suspended material compared to areas where the water is high in suspended material (Chow, 1964). The suspended material blocks the fine pores through which water must pass. Salt solutions in the water affect the viscosity of the fluid and form complexes with the soil colloids, to change the rate of swelling when wet and result in a lower infiltration capacity (Linsley, 1975).

6.3.5 Vegetation

The type and amount of vegetation on the soil has a great influence on infiltration capacity. Vegetal cover increases infiltration compared with barren soil because: 1) it retards surface flow, giving the water additional time to enter the soil; 2) the root systems make the soil more pervious; and 3) the vegetation shields the soil from raindrop impact and reduces compaction by rain (Linsley, 1975).

The vegetative cover also provides a layer of decaying organic matter which promotes the activity of burrowing insects and animals (Wisler, 1967). This tends to increase the pore space available for infiltration and, therefore, increases the infiltration capacity.

6.3.6 Season

The average infiltration capacity of a drainage basin is greatly affected by season. The various factors that affect infiltration capacity are not constant throughout the year and change seasonally (Wisler, 1967). In general infiltration capacity is greatest from mid-August to mid-October, and at a minimum from mid-January to mid-March (Figure 6.1). For coarse sandy soils, there might be considerably less seasonal variation, whereas for a compact clay there might be more seasonal variation (Wisler, 1967).

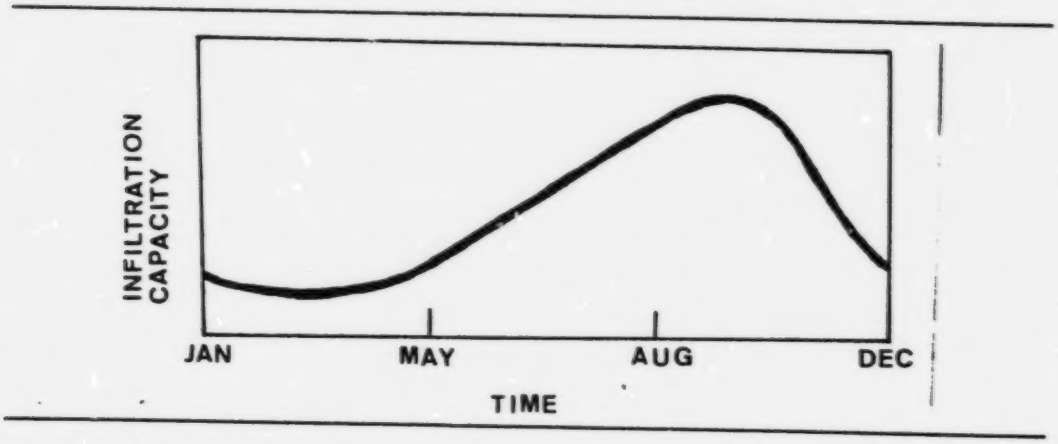

Figure 6.1 Seasonal variation of infiltration capacity (Wisler, 1967).

6.4 INFILTRATION INTO FROZEN SOILS

Infiltration into frozen soils is important in northern latitudes, such as New Brunswick. Unfortunately, only very limited quantitative data are available on the infiltration rates of frozen soils.

The moisture content of a soil at the time of freezing is an important factor which governs the infiltration rate because it determines the quantity and size of ice-free pores (Gray, 1970). If a soil is frozen at a high moisture content, its infiltration rate will be very low, and if saturated, the rate is virtually zero (Figure 6.2). When saturated soil is frozen, an impervious ice layer develops on the surface during the melting period resuting in a very low rate of infiltration. When a soil is frozen at a medium moisture content (60-70% field capacity) the infiltration rate will increase with time (Figure 6.2). In this case, some of the melt water is able to penetrate the soil, and thus transfer heat to melt the ice filled pores and thereby increase the rate of infiltration with time (Gray, 1970). When the soil is frozen at a low moisture content and the soil temperature is near or above freezing at the time of snowmelt, the infiltration rate proceeds as under normal unfrozen conditions (Figure 6.2). In this case only the small pores are filled with ice, and these are thawed rapidly with the downward movement of water (Gray, 1970).

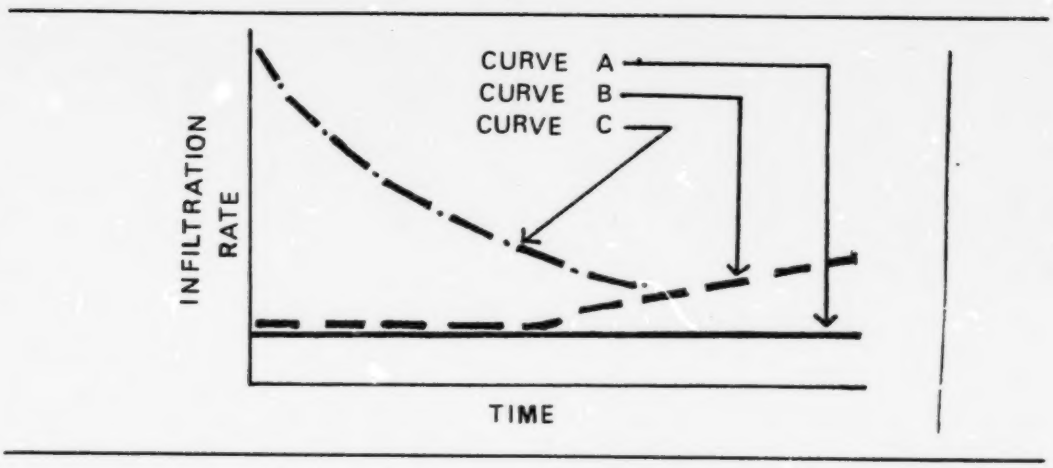

Figure 6.2 Infiltration rate vs time for frozen soils under various soil moisture conditions. Curve A is for a soil frozen at saturated moisture content, Curve B is for a soil frozen at medium moisture content, and Curve C is for a soil frozen at low moisture content.

6.5 MEASUREMENT OF INFILTRATION

There are two general methods of determining infiltration capacity. The first is with an infiltrometer in which water is artificially applied to a small area or sample plot, and the second is by analysis of the hydrograph of runoff resulting from a natural rainfall event on a drainage basin.

The infiltrometer method is helpful in determining the effects of land use, slope, vegetal cover and other variables over which it is desirable to have control in making studies on the prevention of soil erosion, flood reduction and underground storage (Wisler, 1967). The hydrograph analysis is extremely useful for obtaining quantitative data on infiltration capacity (Gray, 1970).

6.5.1 Infiltrometers

Infiltrometers are used on small watersheds or on experimental or sample areas within large watersheds. They are classed as either being a sprinkler or flooding type. The disadvantages of infiltrometers are: 1) the results are representative only of the soil and vegetal conditions of the plot; 2) the natural drops of rainfall cannot be reproduced exactly; and 3) great care must be taken in the installation to reduce boundry effects (Gray, 1970). The results of infiltrometer tests can only be considered as qualatative until they are confirmed by hydrograph analysis because of the artificial conditions in which they are used. However, the data obtained from these tests are useful in defining areas of equal infiltration which can result in a greater precision in analysing the flow conditions of a basin.

6.5.1.1 Flooding Type Infiltrometer

The flooding type infiltrometer is normally one of two types: tubes or conentric rings.

The most common flooding type infiltrometer consists of two shallow concentric rings of sheet metal, usually ranging from nine to thirty-six inches (22 to 91 cm) in diameter (Wisler, 1967). The rings are placed in the ground with the upper portion projecting above the surface. Water is applied, and kept at the same level, in both compartments. Measurements of applied volumes of water are only taken on

the inner ring. The outer ring serves only to reduce the boundary effects on the inner ring (that is lateral flow) (Gray, 1970). From the rate at which water must be added to the inner ring in order to maintain a constant level, the infiltration capacity and its manner of variation are determined.

The tube infiltrometer is usually about nine inches (22 cm) in diameter and 18 to 24 inches (46 to 61 cm) long (Chow, 1964). The tubes are placed deep into the soil with only a few inches (8 cm) above ground surface. Water is poured into the tube and a constant head maintained. The amount of water added at successive time intervals is determined and thus permits direct determination of rates and amounts of infiltration.

The flooding type infiltrometers normally yield values of infiltration about twice those obtained from sprinkler type infiltrometers. Therefore, the results should only be considered to apply under similar fields conditions - for example, flood irrigation (Gray, 1970).

6.5.1.2 Sprinkler Type Infiltrometer

The sprinkler type infiltrometer uses a sprinkler system to simulate rainfall on a test plot. There are many different types of infiltrometers but the most common types are the type F and type FA (Gray, 1970).

The type F infiltrometer generally uses a plot six feet by twelve feet (2 m x 4 m) (Wisler, 1967). Simulated rainfall is applied to the plot and surrounding area from two rows of spray nozzles mounted along each long side of a plot. The nozzles direct their spray upward and slightly inward to cover the plot with relatively uniform rainfall intensities of about 1.75, 3.50, or 5.25 inches per hour (4.4, 8.9 or 13.3 cm/hr), depending on the number of nozzles used (Chow, 1964). The water drops reach a height of about six to seven feet (2 to 2.3 m) above the plot surface.

The FA type rainfall simulator uses a slightly smaller plot than the F type. The normal size of the plot is one foot by two and one-half feet (0.33 to 0.46 m) (Wisler, 1967).

This type of infiltrometer uses the same nozzles as the F type, but they are normally operated at a lower water pressure, and therefore the water does not reach as great a height above the plot surface.

For both the F and FA infiltrometers actual intensity of rain provided by the nozzles is determined by placing a plastic sheet over the plot and by measuring the runoff from the plot (Gray, 1970). After the sheet has been removed, the "test" run is started and continued until the rate of runoff becomes constant. The rainfall simulator is shut off and measurements of discharge continue until runoff ceases. This portion of the test is called the "analytical" run.

Infiltration can then be calculated by the following equation:

$$F = P - S_R - D_A - S_d \qquad \text{(Eq. 6-1)}$$

where F = depth of water infiltrated;
P = depth of rainfall;
S_R = total depth of runoff;
D_A = depth of water on the soil surface; and
S_d = depression storage.

Consistent units of depth must be used to solve the equation. The terms P and S_R are measured quantities for the run, but D_A and S_d must be calculated.

The depth of water on the soil, D_A, is determined from the runoff following the cessation of rain (Gray, 1970). However, there is continuous infiltration during this period which must be accounted for. The infiltration rate during this period may be estimated by the equation:

$$f_r = \frac{f_c}{Q_c} \times Q_r \qquad \text{(Eq. 6-2)}$$

where f_r = infiltration rate at any time during the recession runoff;
f_c = infiltration rate at the cessation of rainfall;
Q_c = the discharge rate at the cessation of rainfall; and
Q_r = the measured discharge rate at any time during the recession.

Each factor has units of depth/hour. The calculated values of f_r are added to the observed values of Q_r and then plotted against time. D_A is then calculated, at any choosen time, as the area under the curve (Chow, 1964).

The depression storage term, S_d, is determined from the "analytical" run. At this time the infiltration rate is at a constant low value equal to the rainfall rate minus the discharge rate at the time when rainfall ceases. S_d may then be calculated by:

$$S_d = P - S_R - D_A - F \qquad \text{(Eq. 6-3)}$$

where: the terms are as above

Since S_d and D_A can be calculated for any discharge outflow from the plot, the run data can now be used with the general equation (Eq. 6-1) to give values of infiltration versus time from the beginning of the test.

6.5.2 Hydrograph Analysis

The curve of infiltration, as it progresses with time, can be estimated for a plot or watershed through analysis of the hydrograph of runoff. This method has advantages over the infiltrometer techniques in that they consider the variability of rainfall and incorporate the influences of factors, such as length of overland flow, slope, soil and vegetation differences, depression storage, and surface

detention as they occur in actual application (Chow, 1964).

As a result of the lag between the time rain falls and the time the excess appears as surface runoff, it is often difficult to determine the exact variation of infiltration capacity. For small basins this lag is often negligible and the actual infiltration capacity can be determined. However, for large basins it is only possible to determine the average infiltration capacity (Wisler, 1967).

6.5.2.1 Determination of f Curves

The infiltration capacity (f) curve can be determined from the hydrograph of runoff by relating the amount of precipitation to the amount of runoff (Figure 6.3).

For small basins, the amount of infiltration may be determined by subtracting the amount of runoff (area under the runoff curve, Figure 6.3) from the amount of precipitation that falls. The value obtained is not a true value of infiltration because it includes interception and depression storage. However, since interception and depression storage never become surface runoff anyway, they may conviently be included as infiltration (Wisler, 1967). To obtain a value of infiltration capacity, it is necessary to divide each value of infiltration by the average length of time during which infiltration was occuring at capacity rates. This length of time is equal to the point of inflection on the recession side of the hydrograph (Chow, 1964).

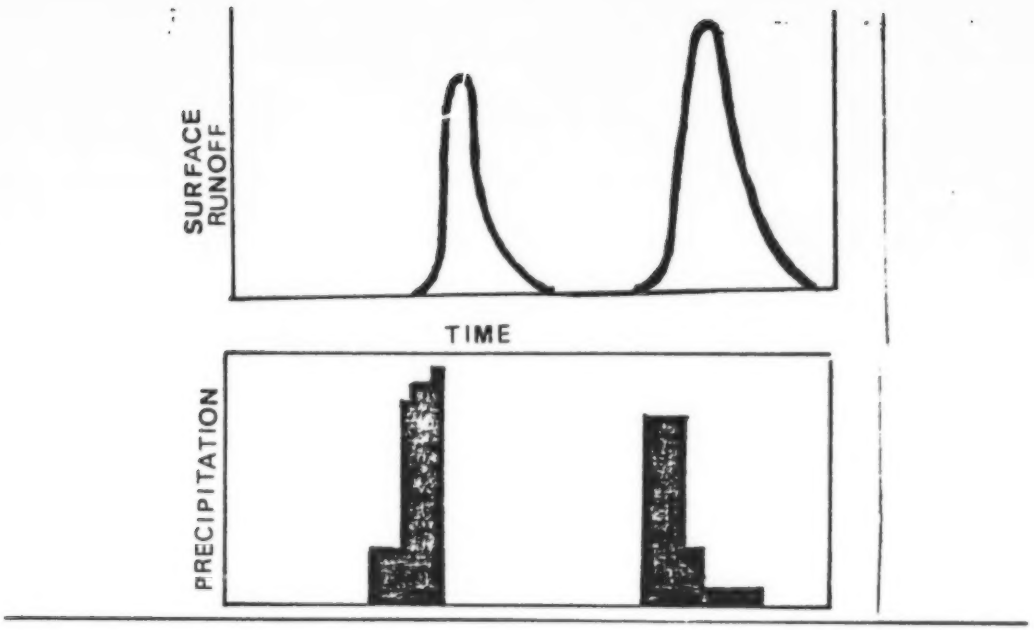

Figure 6.3 Simplified hydrograph of surface runoff and precipitation chart used to calculate infiltration capacity.

For large basins, only an average infiltration capacity curve can be obtained. An assumption is made that the rainfall excess is equal to surface runoff (Wisler, 1967). The rainfall excess is equal to the area of the precipitation graph, above the average infiltration capacity curve. By relating surface runoff to excess rainfall, one can obtain an average infiltration capacity curve by trial and error.

CHAPTER 7
EVAPORATION NETWORKS

7.1 GENERAL

An evaporation network is an integrated system of data collection in which estimates of evaporation are obtained from many stations in a region. The stations in the network are located such that estimates of evaporation can be made in any part of the region by extrapolation. The data obtained from the network are used in developing relationships between evaporation, meteorological factors, and physical parameters, such as slope, elevation, morphology, geology, and soil types (WMO, 1974).

A variety of parameters, including: daily water loss; precipitation; wind speed; and maximum and minimum water and air temperature, are measured at an evaporation station (WMO, 1974). Daily water loss is measured using an evaporation pan of standard national design (U.S. Class A pan for New Brunswick). Precipitation is measured using standard 8 inch rain gauges or tipping bucket rain gauges. Wind speed is measured using a totalizing anemometer and temperature is measured with maximum and minimum thermometers.

Ideally, all the evaporation pans in the network should have the correct exposure, and this is usually achieved by taking care that pans are fully exposed to both wind and

solar radiation (Hounam, 1973). With all the pans in the network under similar conditions of exposure, comparisons between station observations can be made. This allows managers to assess the importance of evaporation in studies of the water budget of catchments (WMO, 1974).

7.2 MINIMUM NETWORK

The first step in a network design is the establishment of a minimum network composed of the minimum number of stations needed to develop relationships between evaporation and the region (WMO, 1974). Where a minimum network is established, general characteristics of evaporation may be determined at any place by extrapolation of values between stations (WMO, 1974). The minimum evaporation network provides data that can be used in studies of spatial and temporal variability of evaporation throughout the region, and provides data that permits the use of evaporation equations.

All the stations in the minimum network are principal or base stations. These stations furnish the basis for statistical studies and thus should be observed continuously and indefinately (WMO, 1974). Although there are no set criteria to determine the density of evaporation stations, the World Meteorological Organization suggests the minimum recommended station density is one station per 100,000 square kilometres over relatively flat terrain (Murray, 1982).

As a result of the low density of stations in the minimum network, it is very important that the records of all stations be of good quality. The type of evaporation pan must be specified since the dimensions and method of exposure have a strong control over the evaporation rate (Hounam, 1972). Therefore, it is necessary to label all data with equipment details. Once the minimum network has been established it is important to work towards establishing an optimum network.

7.3 OPTIMUM NETWORK

An optimum evaporation network consists of many stations widely distributed throughout a region. The aim of the optimum network is to make it possible, by interpolation between stations, to determine, with sufficient accuracy for practical purposes, all quantitative data that define the statistical distribution of evaporation at any location in the region (WMO, 1974).

There are three types of stations in an optimum network: Base stations; secondary stations; and specialized stations. The base or principal stations are the same as those in the minimum network. Secondary stations are used to establish good correlations between them and the base station or with characteristics of the terrain and therefore are only operated for a limited number of years (WMO, 1974). The secondary stations are moved to new locations once

correlations have been established. Specialized stations are used to augment the network of base and secondary stations. The length of observation of a specialized station is determined by the purpose for which they were installed. In some cases, the observations may be confined to one particular aspect or to one season of the year. Stations that measure evapotranspiration, such as lysimeter stations, are specialized stations.

There are not any universal guidelines that are used to determine the number of stations needed for an optimum network. However, enough stations should be established to take into account climatic and topagraphic variability, with denser networks where the variabilities are greater (Atmospheric Environment Service, 1978). The establishment of an optimum evaporation network is essential in arid regions in which the water lost by evaporation represents a significant amount in the water balance of a drainage area.

7.4 EVAPORATION NETWORK FOR NEW BRUNSWICK

7.4.1 Existing Network

Currently, evaporation data is collected from two stations in New Brunswick. These stations are located at Fredericton CDA and Mount Carleton. A third station located at CFB Chatham, is due to begin operation May 1, 1987. In addition to evaporation, temperature, precipitation, rate of

rainfall, wind speed, sunshine and snow depth are recorded at these locations (Murray, 1982).

7.4.2 Minimum Network

New Brunswick has an area 73,405 km^2 and therefore, according to the World Meteorological Organization guidelines, needs only one station to meet the minimum aerial standard for a relatively flat area. Since the network density in New Brunswick is one station per 24,801 km^2, a satisfactory minimum network exists.

The existing stations represent a satisfactory minimum network, but the data obtained may not be adequate to estimate evaporation throughout the Province. A network of pan evaporiometers is a basic requirement as an aid to the development and evaluation of empirical formulae (Hounam, 1972).

7.4.3 Secondary Stations

Additional secondary stations should be established throughout the province in order to collect the data required for development of empirical equations. The additional stations should be located so that the variability in physiography, climate, geology and hydrology is sampled. An analysis of the physiographic, hydrogeologic and hydroclimatic regions can be done to define major divisions of the province.

New Brunswick has five distinct physiographic regions (Figure 7.1). These regions include: the Gulf of St. Lawrence Plain; the N.B. lowland; the N.B. Highland; the Chaleur upland; and the St. Croix/Caledonia Highlands. Each region is composed of different rock types and has a different average elevation.

New Brunswick also has four major hydroclimatic regions (Figure 7.2). These include the Northern Uplands, the Coastal lowlands, the Coastal Flatlands and the Southern Highlands.

Finally five hydrogeological regions are present (Figure 7.3). These regions are deliniated based on topography, soil type, geologic origin and weather regimes.

When the physiographic, hydroclimatic and hydrogeologic regions of the Province are compared, it is evident that five distinctly different regions exist in New Brunswick (Figure 7.4). Therefore, it is suggested that as a start to a good optimum network, one secondary station should be established in each major region.

The density of the population and economic activity will influence the location of the secondary station within each region (WMO, 1974). A variety of possible locations for each region are given in Table 7.1.

7.4.4 Specialized Stations

In addition to establishing secondary stations, some specialized stations might also be operated. These stations should be located near major reservoirs or large lakes, or in areas where the demand for water is great. Specialized stations are normally operated by an agency for their own needs. The data they provide, however, may be of interest to other organizations. Possible locations for specialized stations are given in Table 7.1.

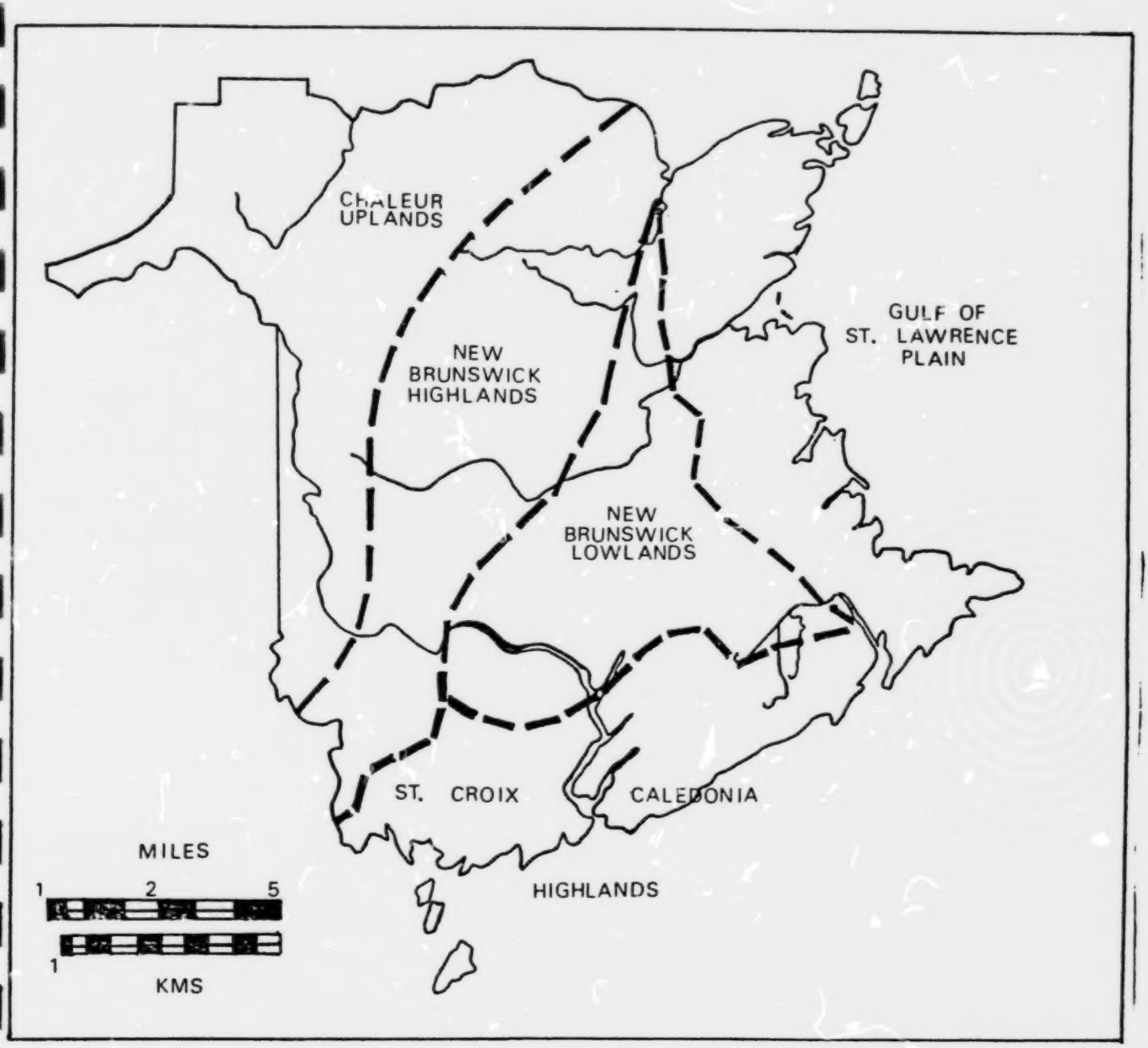

Figure 7.1. The physiographic regions of New Brunswick (Environment Canada, 1974).

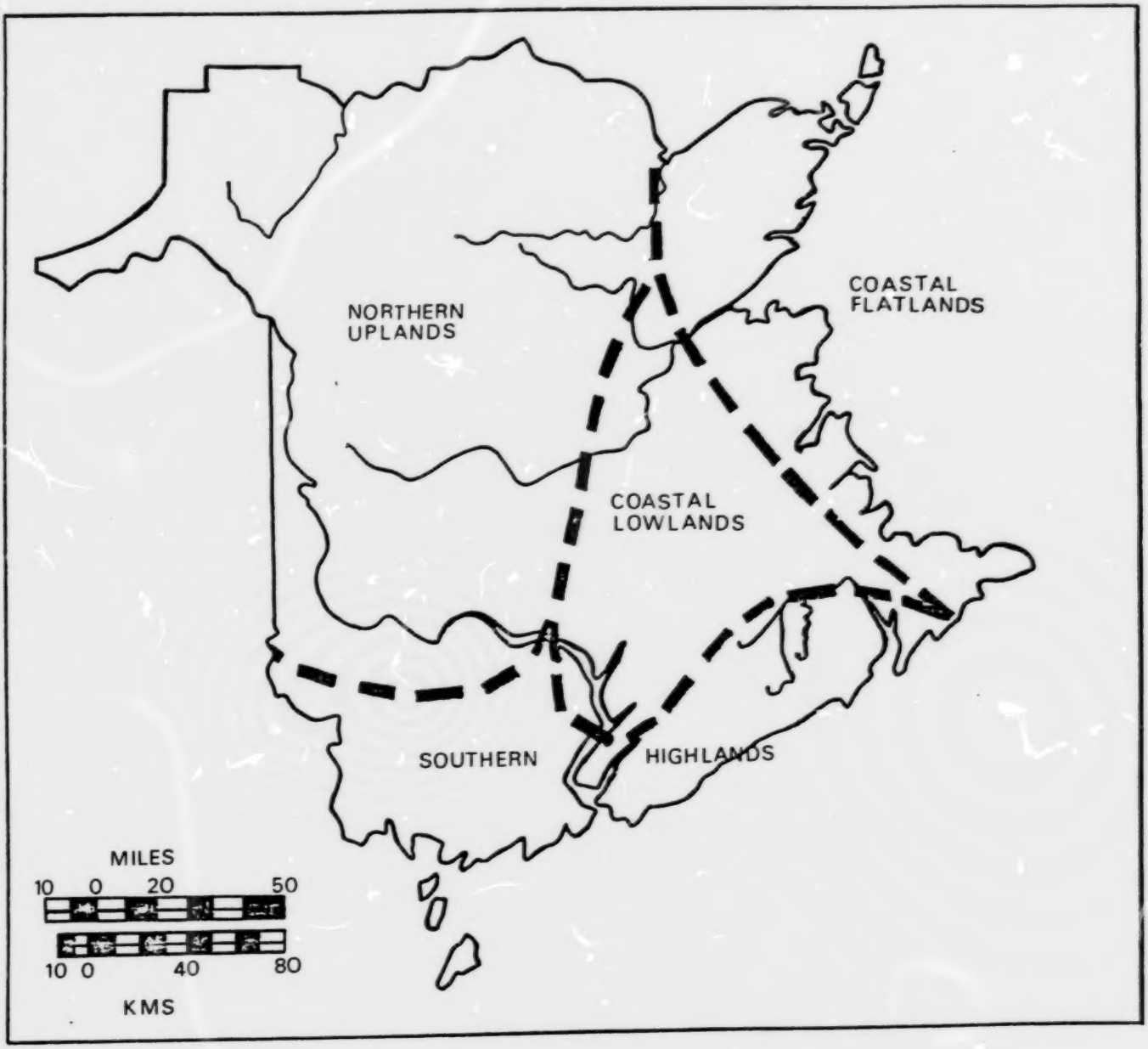

Figure 7.2. The major hydroclimatic regions of New Brunswick (Environment Canada, 1978).

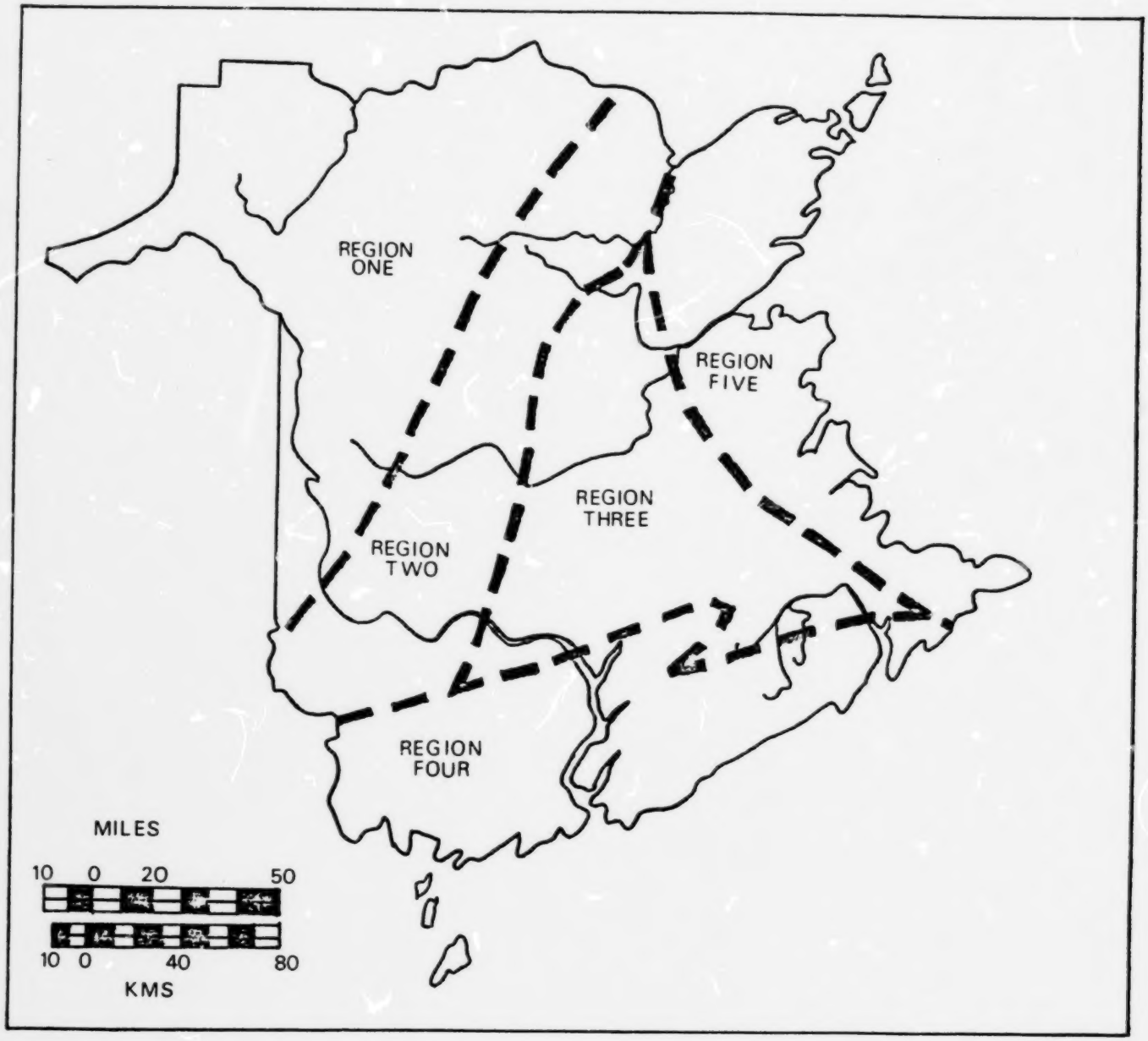

Figure 7.3. The hydrogeological regions of New Brunswick. The regions are deliniated based on topography, soil type, and geologic origin of the rocks. (Canada Flood Damage Program, 1977)

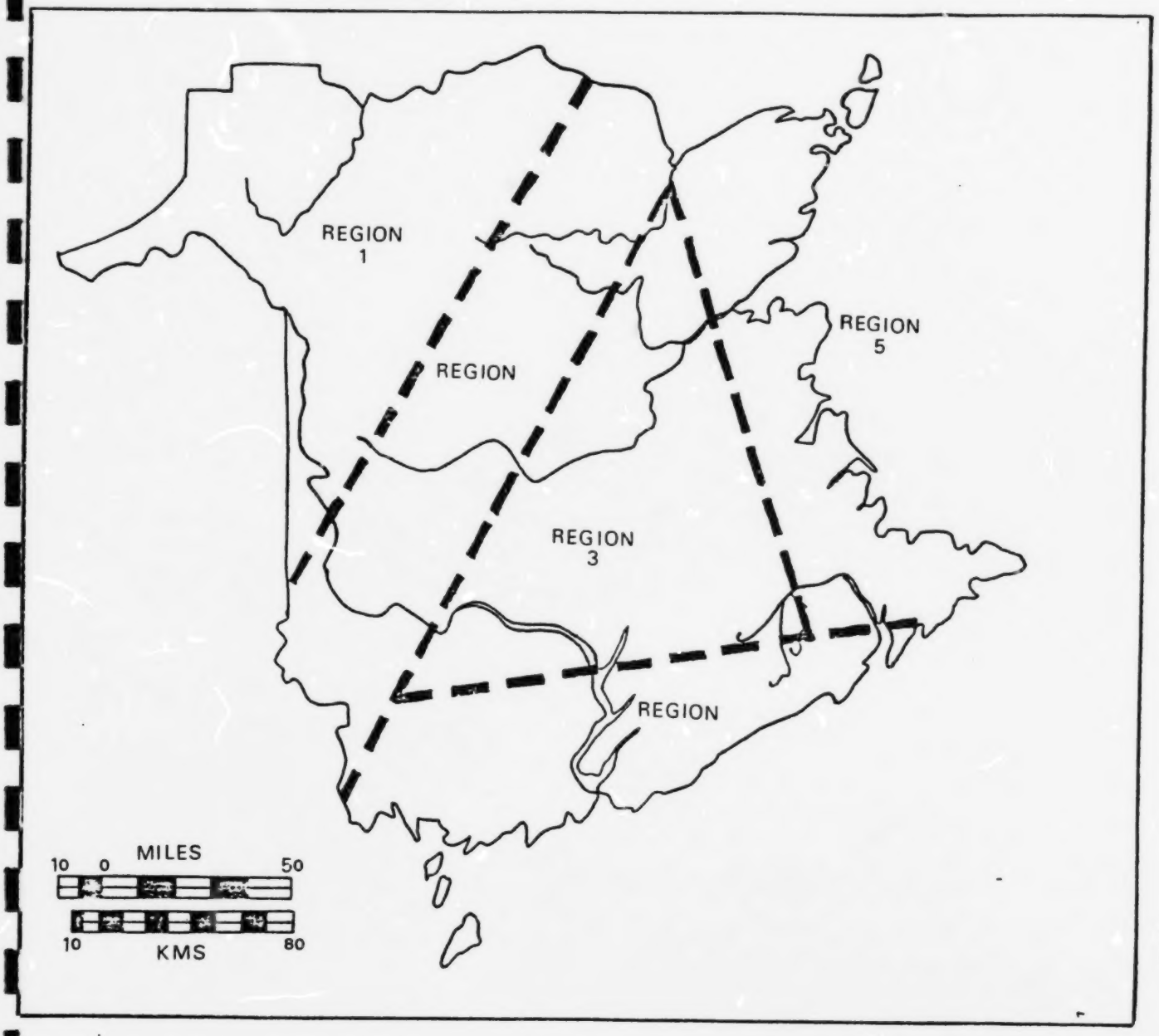

Figure 7.4. Geographic divisions of New Brunswick based on physiographic and hydroclimatic regions.

Table 7.1. Possible locations for principal, secondary and specialized Stations for New Brunswick.

	REGION	POSSIBLE LOCATION	COMMENTS	ALTERNATE LOCATION
Principal Station	3	Fredericton CDA	near boundary of three regions. Existing Station present	-----
Secondary Stations	1	Mount Carleton	Existing Station; highest elevation.	Grand Falls
	2	Bathurst	Major population centre.	Chatham
	3	Boiestown	Major forestry centre.	Chipman
	4	Alma	Tourist centre Agriculture centre.	Sussex
	5	Chatham	Existing station	-----
Specialized Stations	1	Beechwood	Major reservoir	-----
	2	Mactaquac	Major reservoir	-----
	3	Grand Lake	Large lake; tourist area	----- -----
	4	Turtle Creek	Major reservoir for Moncton	-----

CHAPTER 8
CONCLUDING REMARKS

Interception, evaporation, transpiration, evapotranspiration and infiltration are components of the hydrologic cycle, important with regard to the establishment of water balances for watersheds. The hydrologist must have an understanding of these components in order to evaluate their influence.

There are a wide variety of instruments available to measure each component in the hydrologic cycle. These instruments include evaporation pans, lysimeters and atmometers. The instrument chosen will be dependent on the nature of the results required; be it annual estimates of evaporation, rates of infiltration, or hourly estimates of potential evapotranspiration.

For practical considerations, it is highly desirable to be able to obtain estimates of evaporation and evapotranspiration from climatological means. This may be accomplished with the use of empirical equations. However, in order to use these equations empirical coefficients must be defined.

In New Brunswick further study needs to be done to develop better equations for the estimation of evaporation and evapotranspiration from climatological data. The development of reliable and accurate equations will eventually reduce the need for direct measurements of evaporation and evapotranspiration.

However, until these equations are developed, there is still a need to collect evaporation data.

Data collection is essential for the development of better equations. The Class A evaporation network in New Brunswick should be expanded to provide data that can be used to derive localized coefficients for equations. The stations in the network should be located such that the climatic and hydrologic variability across the province is considered.

Interception and transpiration, although significant components of the hydrologic cycle, are difficult to quantitatively measure. In many cases, their contribution to the water balance of a region is not fully understood and thus furthur research on these topics should be done.

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APPENDICIES

APPENDIX I

TABLES AND FIGURES USED FOR COMPUTATION
OF EVAPORATION AND EVAPOTRANSPIRATION
EQUATIONS

LIST OF TABLES AND FIGURES

Table a. Saturation vapor pressure (e_s) in inches of Mercury.

Table b. Mid-monthly intensity of solar radiation on a horozontal surface in mm water per day, or the value R in the Penman equation.

Table c. Mean possible duration expressed in units of 30 days of 12 hours each, or the adjusting factor for potential evapotranspiration computed by the Thornthwaite equation.

Table d. Value of σT_a^4 corresponding to temperature in degrees farenheit.

Figure a. Slope of saturated vapor pressure curve for air temperature.

Saturation Vapor Pressure (e_s) in Inches of Mercury
(After Isler)

Temperature °F	e_s	Temperature °F	e_s	Temperature °F	e_s	Temperature °F	e_s
-30	0.007	10	0.063	50	0.363	90	1.423
-29	.007	11	.067	51	.376	91	1.469
-28	.008	12	.070	52	.390	92	1.515
-27	.008	13	.074	53	.405	93	1.563
-26	.009	14	.077	54	.420	94	1.612
-25	.010	15	.081	55	.436	95	1.662
-24	.010	16	.085	56	.452	96	1.714
-23	.011	17	.089	57	.469	97	1.767
-22	.011	18	.094	58	.485	98	1.822
-21	.012	19	.099	59	.504	99	1.878
-20	.013	20	.103	60	.522	100	1.936
-19	.013	21	.108	61	.541	101	1.994
-18	.014	22	.114	62	.560	102	2.055
-17	.015	23	.119	63	.580	103	2.117
-16	.016	24	.125	64	.601	104	2.181
-15	.017	25	.131	65	.623	105	2.246
-14	.018	26	.137	66	.645	106	2.314
-13	.019	27	.143	67	.668	107	2.382
-12	.020	28	.150	68	.691	108	2.453
-11	.021	29	.157	69	.715	109	2.525
-10	.022	30	.165	70	.740	110	2.599
-9	.024	31	.172	71	.766	111	2.676
-8	.025	32	.180	72	.792	112	2.754
-7	.026	33	.188	73	.819	113	2.833
-6	.028	34	.195	74	.847	114	2.915
-5	.029	35	.203	75	.876	115	2.999
-4	.031	36	.212	76	.906	116	3.085
-3	.033	37	.220	77	.936	117	3.173
-2	.034	38	.229	78	.968	118	3.264
-1	.036	39	.238	79	1.000	119	3.356
0	.038	40	.248	80	1.033	120	3.451
1	.040	41	.258	81	1.068	121	3.548
2	.042	42	.268	82	1.103	122	3.647
3	.044	43	.278	83	1.139	123	3.749
4	.047	44	.289	84	1.176	124	3.853
5	.049	45	.300	85	1.215	125	3.960
6	.052	46	.312	86	1.254	126	4.069
7	.055	47	.324	87	1.295	127	4.181
8	.057	48	.336	88	1.336	128	4.295
9	.060	49	.349	89	1.379	129	4.412

Table a

Midmonthly Intensity of Solar Radiation on a Horizontal Surface in mm. Water per Day, or the Value R in the Penman Equation (after Criddle, 1958)

N. latitude deg	J	F	M	A	M	J	J	A	S	O	N	D
90	-	-	-	7.9	14.9	18.1	16.8	11.2	2.6	-	-	-
80	-	-	1.8	7.8	14.6	17.8	16.5	10.6	4.0	0.2	-	-
70	-	1.1	4.3	9.1	13.6	17.0	15.8	11.4	6.8	2.4	0.1	-
60	1.3	3.5	6.8	11.1	14.6	16.5	15.7	12.7	8.5	4.7	1.9	0.9
50	3.6	5.9	9.1	12.7	15.4	16.7	16.1	13.9	10.5	7.1	4.3	3.0
40	6.0	8.3	11.0	13.9	15.9	16.7	16.3	14.8	12.2	9.3	6.7	5.5
30	8.5	10.5	12.7	14.8	16.0	16.5	16.2	15.3	13.5	11.3	9.1	7.9
20	10.8	12.3	13.9	15.2	15.7	15.8	15.7	15.3	14.4	12.9	11.2	10.3
10	12.8	13.9	14.8	15.2	15.0	14.8	14.8	15.0	14.9	14.1	13.1	12.4
0	14.5	15.0	15.2	14.7	13.9	13.4	13.5	14.2	14.9	15.0	14.6	14.3

Table 5

Mean Possible Duration Expressed in Units of 30 Days of 12 Hr Each, or the Adjusting Factor for Potential Evapotranspiration Computed by the Thornthwaite Equation (After Criddle, 1958)

N. latitude deg	J	F	M	A	M	J	J	A	S	O	N	D
0	1.04	0.94	1.04	1.01	1.04	1.01	1.04	1.04	1.01	1.04	1.01	1.04
10	1.00	0.91	1.03	1.03	1.08	1.06	1.08	1.07	1.02	1.02	0.98	0.99
20	0.95	0.90	1.03	1.05	1.13	1.11	1.14	1.11	1.02	1.00	0.93	0.94
30	0.90	0.87	1.03	1.08	1.18	1.17	1.20	1.14	1.03	0.98	0.89	0.88
35	0.87	0.85	1.03	1.09	1.21	1.21	1.23	1.16	1.03	0.97	0.86	0.85
40	0.84	0.83	1.03	1.11	1.24	1.25	1.27	1.18	1.04	0.96	0.83	0.81
45	0.80	0.81	1.02	1.13	1.28	1.29	1.31	1.21	1.04	0.94	0.79	0.75
50	0.74	0.78	1.02	1.15	1.33	1.36	1.37	1.25	1.06	0.92	0.76	0.70

Table c

Temperature	σT_a^4
°F	mm. H_2O/day
35	11.48
40	11.96
45	12.45
50	12.49
55	13.45
60	13.96
65	14.52
70	15.10
75	15.65
80	16.25
85	16.85
90	17.46
95	18.10
100	18.80

Table d

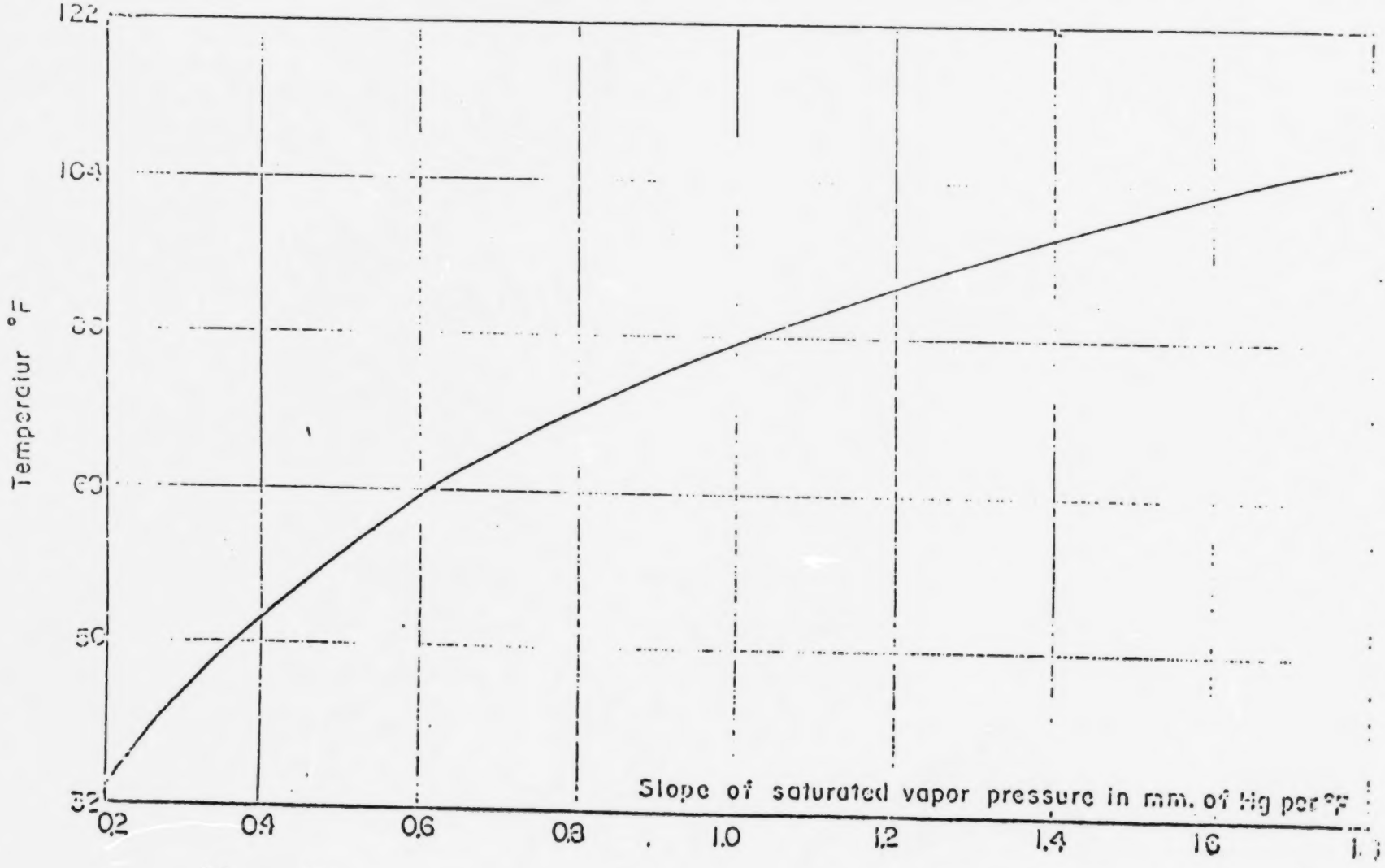


FIG. a SLOPE OF SATURATED VAPOR PRESSURE CURVE FOR GIVEN AIR TEMPERATURE IN (Vapor Pressure)

APPENDIX II

POTENTIAL EVAPOTRANSPIRATION MAPS
OF NEW BRUNSWICK

The following are a series of maps showing mean monthly and total annual potential evapotranspiration in New Brunswick. The data were obtained by analysis of the climatic water balance based on the period 1941-1970. Forty stations were used in the analysis. The iso-lines produced are in increments of 10mm of water. Various point values of potential evapotranspiration are also plotted.

Figure a. Mean monthly potential evapotranspiration in New Brunswick for the month of April.

Figure b. Mean monthly potential evapotranspiration in New Brunswick for the month of May.

Figure c. Mean monthly potential evapotranspiration in New Brunswick for the month of June.

Figure d. Mean monthly potential evapotranspiration in New Brunswick for the month of July.

Figure e. Mean monthly potential evapotranspiration in New Brunswick for the month of August.

Figure f. Mean monthly potential evapotranspiration in New Brunswick for the month of September.

Figure g. Mean monthly potential evapotranspiration in New Brunswick for the month of October.

Figure h. Total annual potential evapotranspiration in New Brunswick.

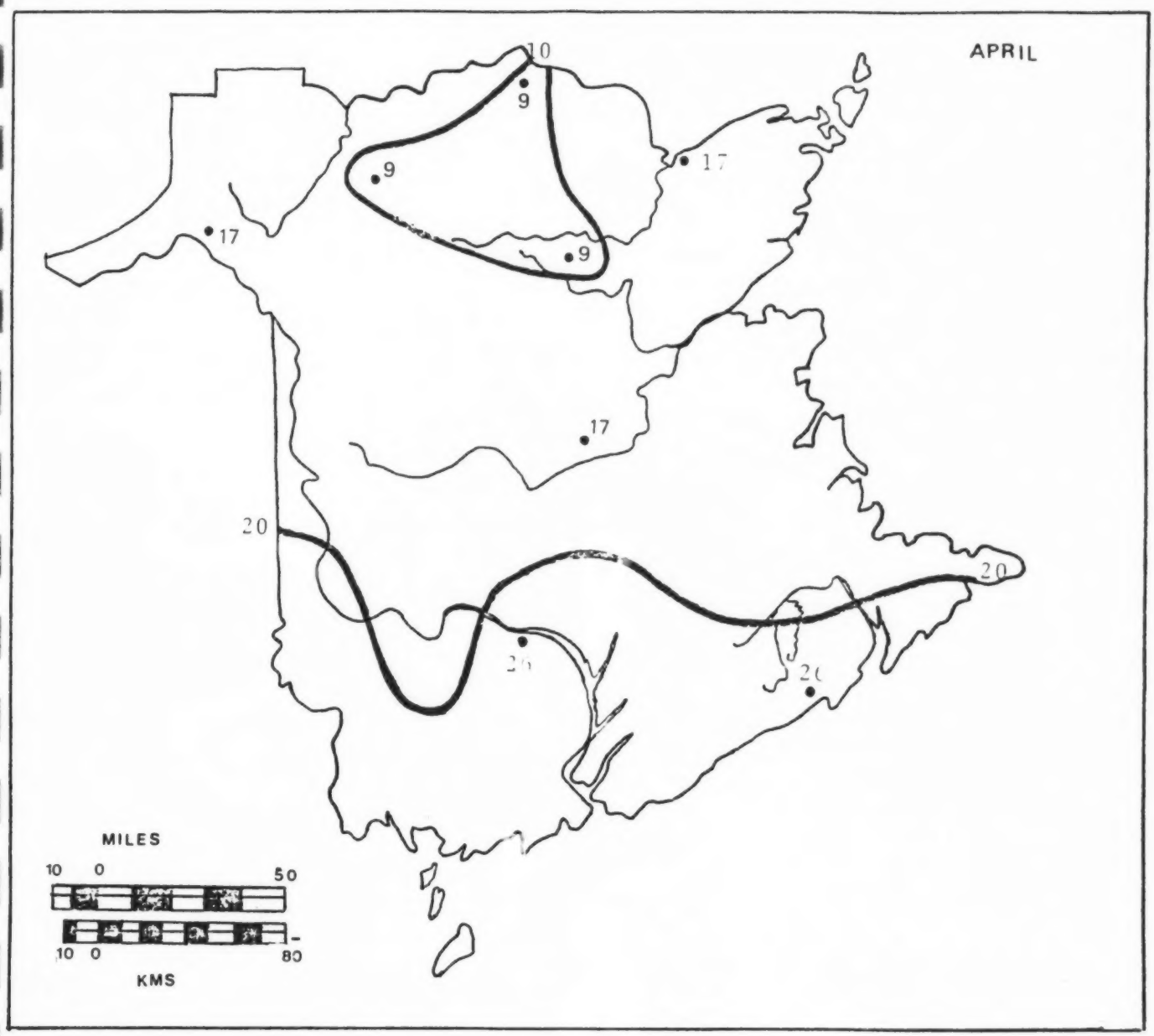


Figure a. Mean monthly potential evapotranspiration in New Brunswick for the month of April. Data are based on the period 1941-1970. Values are in mm of water.

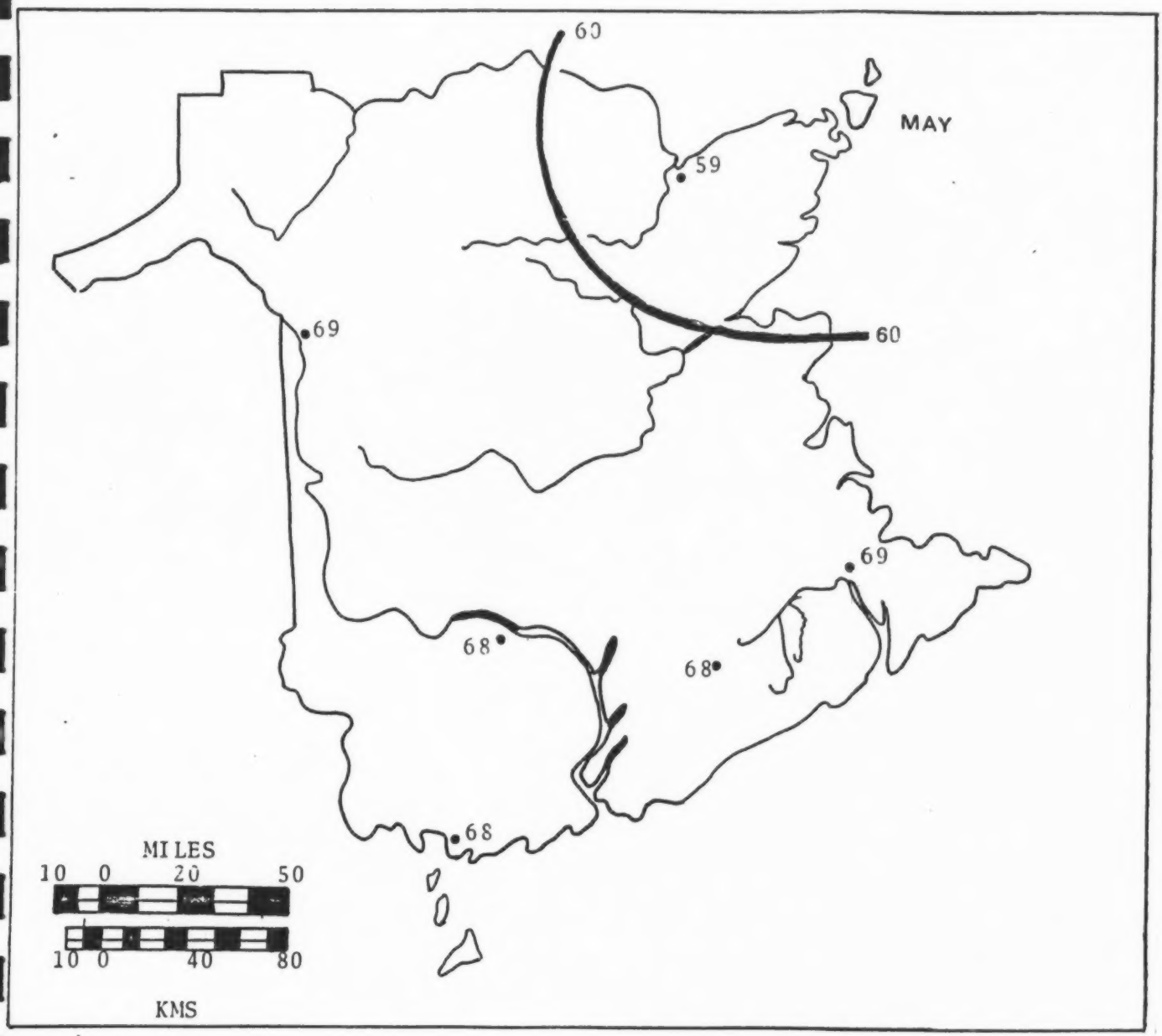


Figure b. Mean monthly potential evapotranspiration in New Brunswick for the month of May. Data are based on the period 1941-1970. Values are in mm of water.

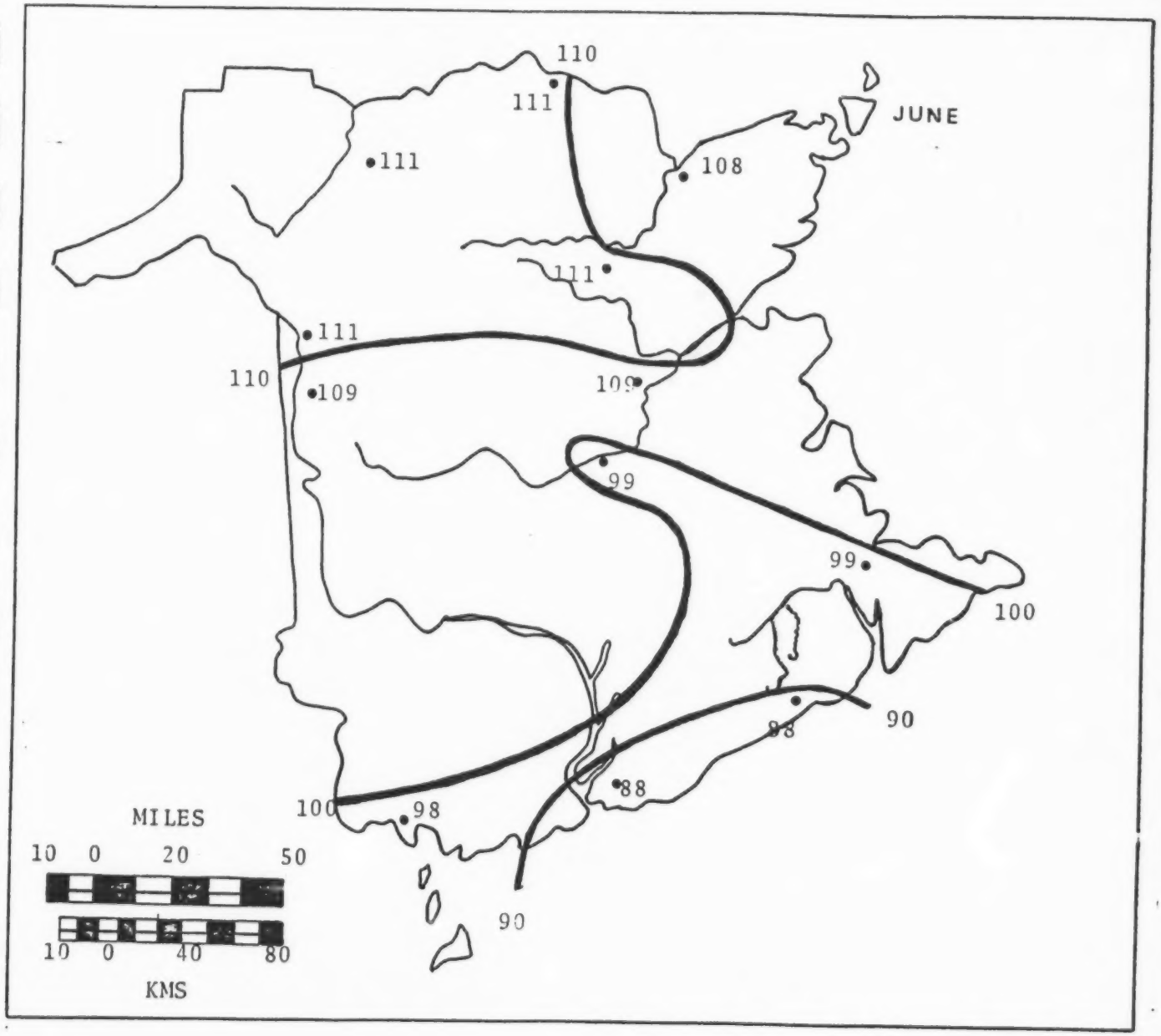


Figure c. Mean monthly potential evapotranspiration in New Brunswick for the month of June. Data are based on the period 1941-1970. Values are in mm of water.

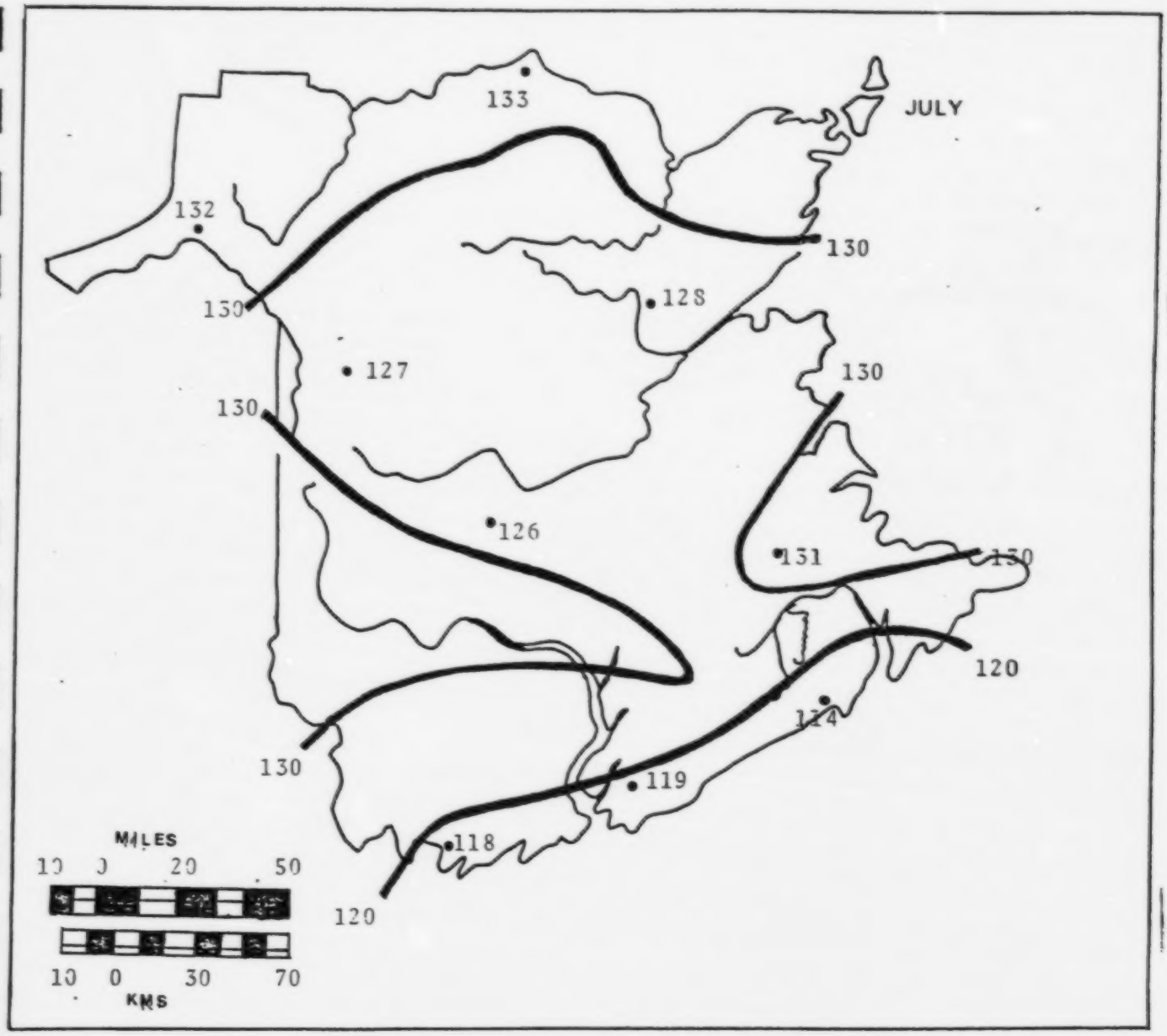


Figure d. Mean monthly potential evapotranspiration in New Brunswick for the month of July. Data are based on the period 1941-1970. Values are in mm of water.

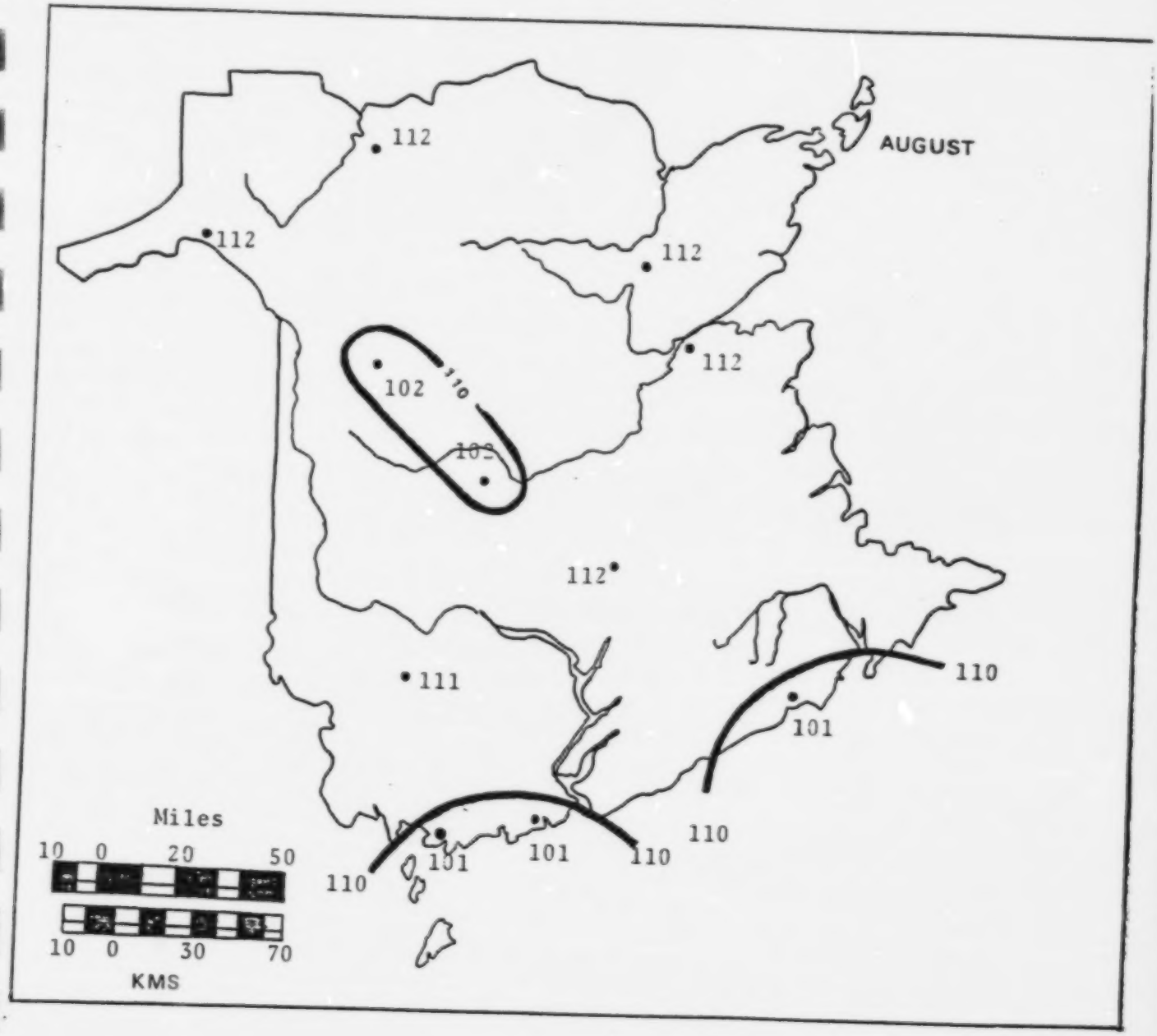


Figure e. Mean monthly potential evapotranspiration in New Brunswick for the month of August. Data are based on the period 1941-1970. Values are in mm of water.

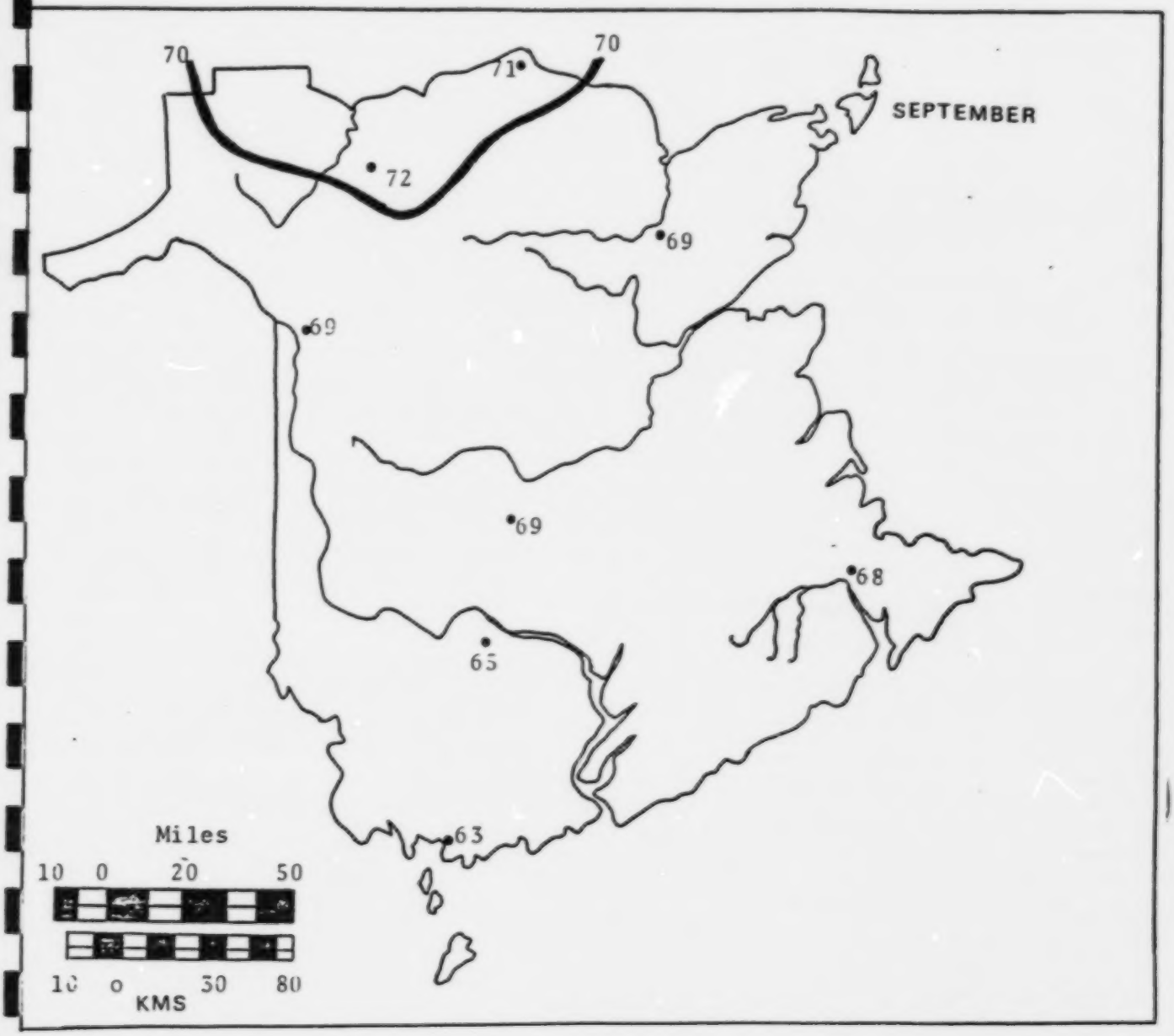


Figure f. Mean monthly potential evapotranspiration in New Brunswick for the month of September. Data are based on the period 1941-1970. Values are in mm of water.

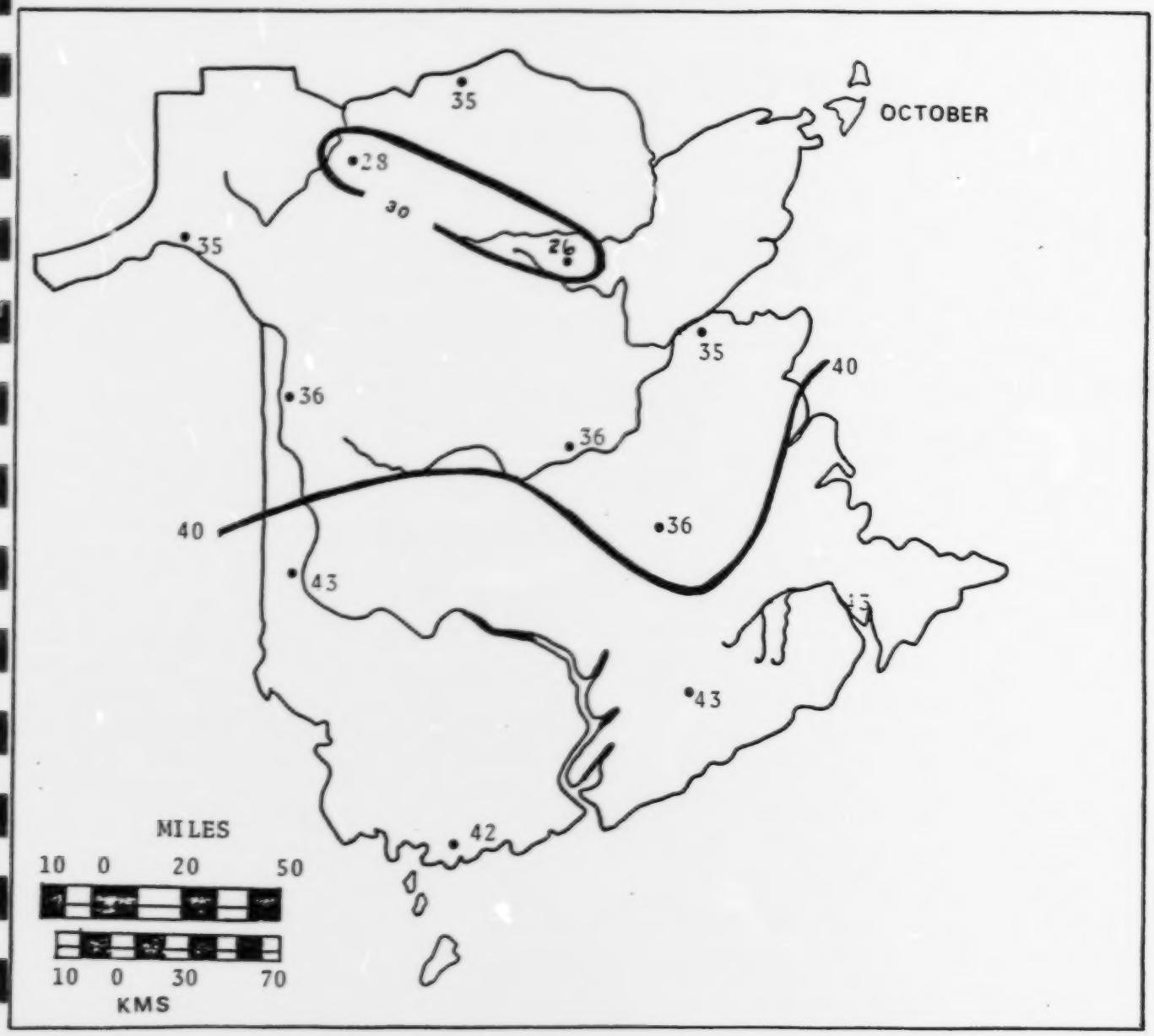


Figure g. Mean monthly potential evapotranspiration in New Brunswick for the month of October. Data are based on the period 1941-1970. Values are in mm of water.

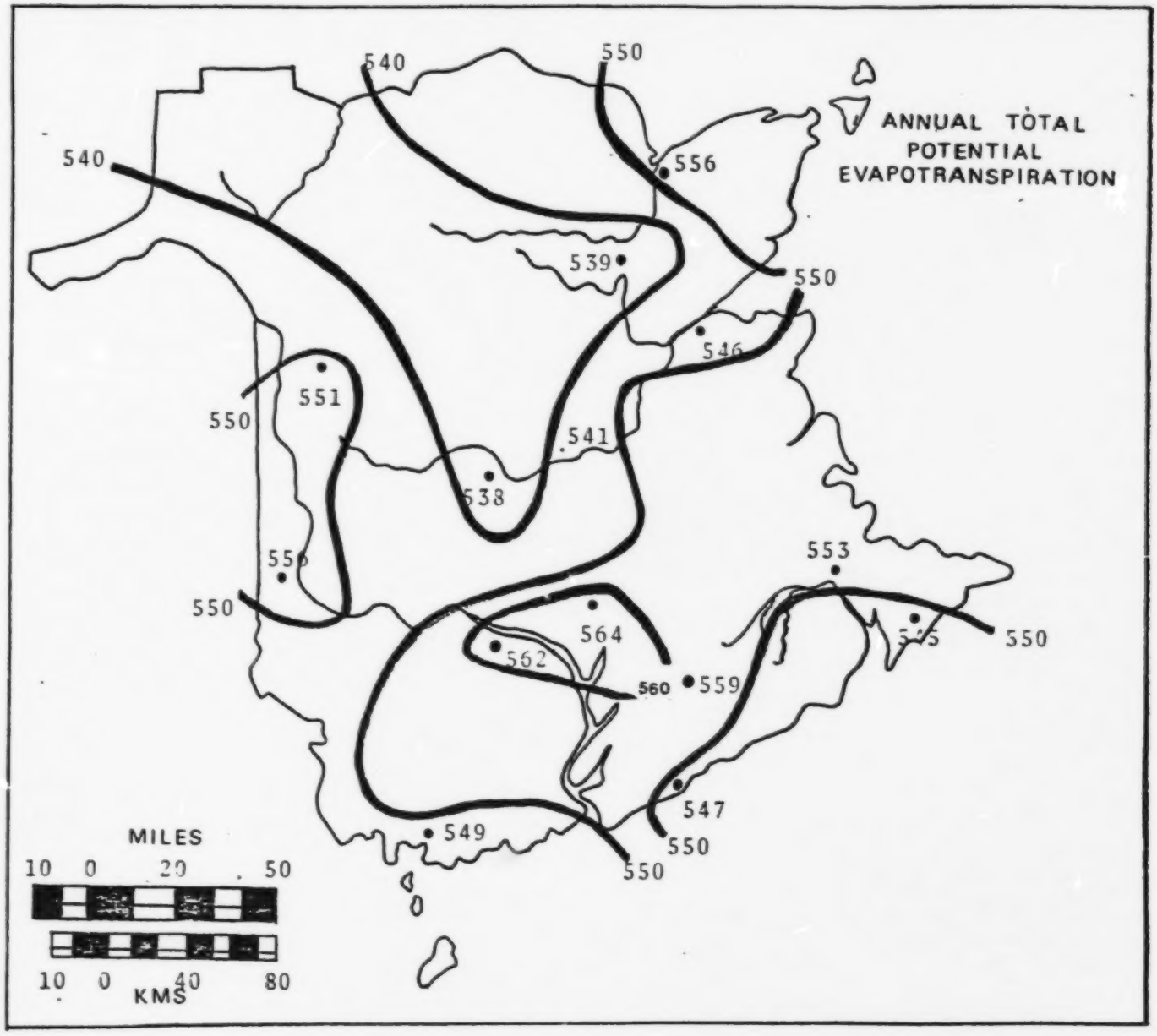
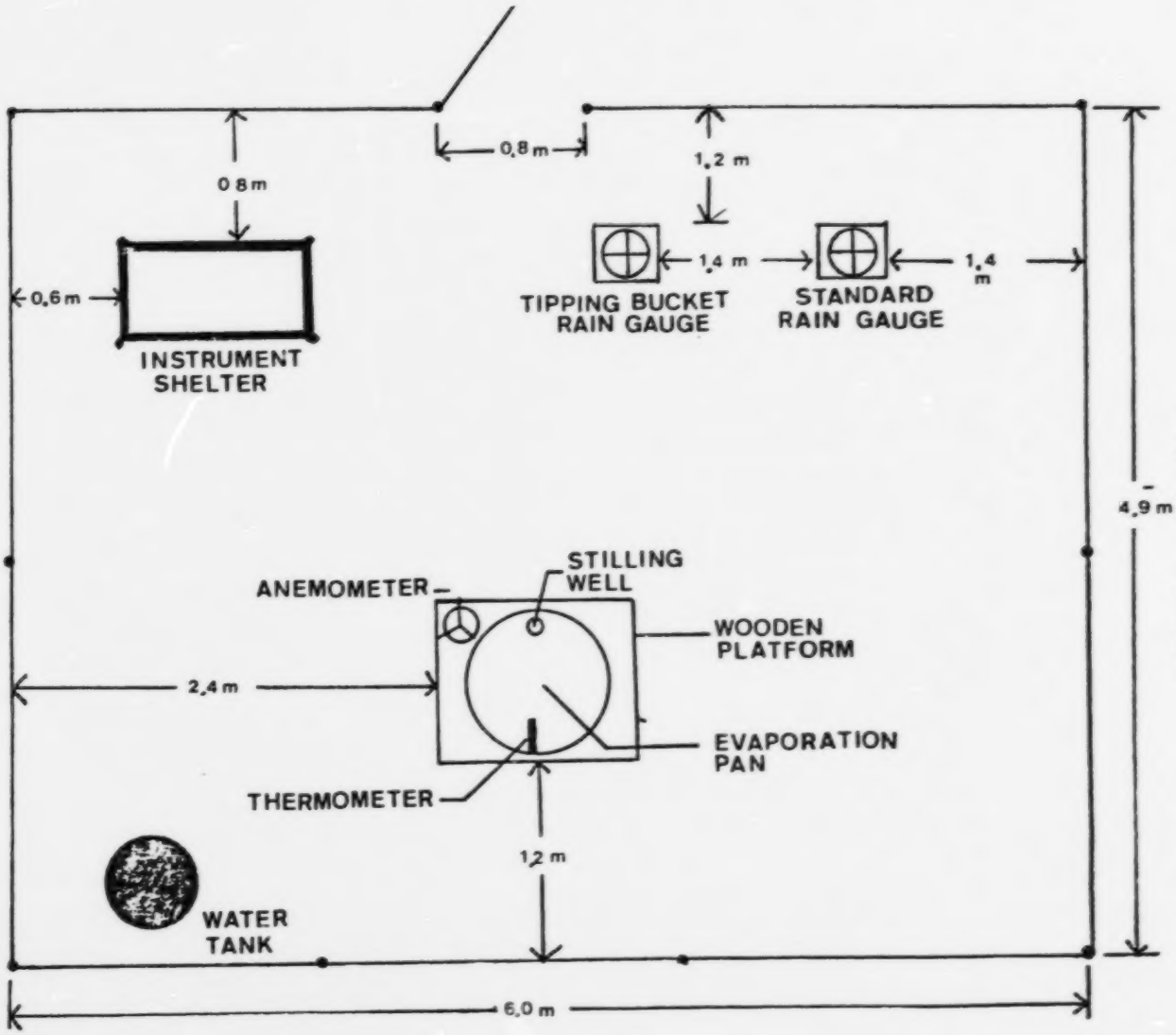


Figure h. Total annual potential evapotranspiration in New Brunswick. Data are based on the period 1941-1970. Values are in mm of water.

APPENDIX III

DIAGRAM OF AN EVAPORATION STATION

The following is a diagram of an evaporation station. Dimensions are given according to AES guidelines.



APPENDIX IV

LIST OF MANUFACTURERS AND DISTRIBUTERS OF EVAPORATION AND EVAPOTRANSPIRATION MEASUREMENT EQUIPMENT

The following are lists of some North American manufacturers and distributers of instruments needed for measurement or estimation of evaporation and evapotranspiration.

LIST OF MANUFACTURERS

1. Belfort Instrument Company
   1600 South Clinton Street
   Baltimore, Maryland
   21224

2. Cole Parmer Instrument Company
   7425 North Oak Park Avenue
   Chicago, Illinois
   60648

3. Enviro-Labs Incorporated
   626 Sonora Avenue
   Glendale, California
   91201

4. Hollis Geosystems Corporation
   18 Clinton Drive
   Hollis, New Hampshire
   03049

5. Sierra - Misco Incorporated
   1825 Eastshore Highway
   Berkeley, California
   94710

6. Leupold and Stevens Incorporated
P.O. Box 688
Beaverton, Oregon
97005

7. Teledyne Gurley Limited
513 Fulton Street
Troy, New York
12181

8. Weathermeasure Corporation
Division of Systron Donner
P.O. Box 41257
Sacramento, California
95841

9. Weathertronics
Division of Qualimetrics Incorporated
2777 Del Monte Street
P.O. Box 1437
West Sacramento, California
95691

LIST OF DISTRIBUTERS

1. Baytronix Limited
   4006 Cote Vertu
   Montreal, Quebec
   H4R 1V4

   74 Main Street North, Suite 5
   Brampton, Ontario
   L6V 1N7

   1300 Carling Avenue, Suite 308
   Ottawa, Ontario
   K1Z 7L2

2. C.J. Munn Equipment Limited
   P.O. Box 3172
   Postal Station "B"
   Fredericton, New Brunswick
   E3A 5G9

3. Comeau Technique Limited
   5708 Vanden Abeele
   St. Laurent, Quebec
   H4S 1R9

4. Eirtech Instruments Incorporated
   1400 Dewey Avenue
   Rochester, New York
   14613

5. Frederick Goertz Limited
   1110 Sheppard Avenue East, Suite 505-1
   Willowdale, Ontario
   M2K 2W2

6. Geneq Incorporated
   7978 Est, Rue Jarry
   Anjou, Quebec
   H1J 1H5

7. Hoskin Scientific Eastern
   1505 Barrington Street, Suite 1205
   Halifax, Nova Scotia
   B3J 3K5

8. Hydrometeo Incorporated
5505 Iberville, Suite 200
Montreal, Quebec
H2G 2B2

9. MIC Company
Division of Hohener Enterprises Limited
Meteorological Instrument Centre
37 Maple Avenue
Thornhill, Ontario
L3T 3S8

10. Science Associates Incorporated
230 Nassau Street
Box 230
Princeton, New Jersey
08540

APPENDIX V

1986 PRICE LIST OF INSTRUMENTS

The following is a 1986 price list for various instruments that might be used when measuring evaporation or evapotranspiration. The information was obtained from distributers throughout Canada.

PRICE LIST

Class A Evaporation Pan	$ 560 -	820
Hook Gauge	$ 280 -	310
Stilling Well	$ 230 -	260
Totalizing Anemometer	$ 540 -	1165
Hand-held Anemometer	$ 210 -	290
Digital / Thermo Anemometer	$ 1110 -	1250
Standard 8 in Rain Gauge	$ 315 -	390
Tipping Bucket Rain Gauge	$ 620 -	1630
Floating Maximum / Minimum Thermometer	$ 150 -	200
Max / Min Air Thermometer	$ 45 -	50
Wet / Dry Bulb Hygrometer	$ 35 -	60
Hygrothermograph	$ 900 -	950
Sling Psychrometer	$ 70 -	80
Aneroid Barograph	$ 850 -	1000
Piche Evaporimeter	$ 90 -	180
Stand For Evaporimeter	$ 95 -	150
Filter Papers For Evaporimeter	$ 85 -	90

Meteorograph	$	1155 - 1270
Charts For Meteograph	$	90 - 120
Campbell Stokes Sunshine Recorder	$	1990 - 3526
Charts For Sunshine Recorder	$	150 - 215
Net Radiometer	$	1120 - 1190
Mast To Mount Radiometer	$	165 - 180
Mechanical Pyranograph	$	1240 - 1280
Mast For Pyranograph	$	160 - 180
Spare Kit For Pyranograph	$	350 - 370
Large Tube Solarimeter	$	1375 - 1430
Instrument Shelter	$	680 - 705
Automatic Weather Station	$	15,400 - 16,000

APPENDIX VI

NEW BRUNSWICK SUPPLIERS OF MATERIAL USED TO CONSTRUCT MEASUREMENT EQUIPMENT

The following is a list of New Brunswick companies who might be able to supply the materials needed to construct various instruments for measurement of evaporation and evapotranspiration

1. A and P Concrete Products Ltd
   Sainte-Marie- De-Kent
   EOA 3AO

2. Atlantic Iron Works Ltd
   P.O. Box 180
   Shediac Bridge
   EOA 3HO

3. Atlas Alloys
   P.O. Box 3607 Stn B
   Saint John
   E2M 4Y2

4. Branscombe's Foundry
   RR #2 Site 5
   P.O. Box 2
   Hampton
   EOG 1ZO

5. Brunswick Fiberglass Ltd.
   P.O. Box 781
   Saint John
   E2L 4B3

6. Campbellton Welding and Machinr Shop Ltd.
   P.O. Box 1022
   Campbellton
   E3N 3H4

7. Canada Cement Lafarge Ltd.
   Havelock
   EOA 1WO

8. Creative Fiberglass Ltd.
   RR #2
   Fredericton
   E3B 4X3

9. Delco Enterprises Ltd.
   P.O. Box 101
   St. Antoine
   EOA 2XO

10. Diamond Concrete Ltd.
P.O. Box 117
Minto
E0E 1J0

11. Downeast Plastics Ltd.
P.O. Box 509
Cap Pele
E0A 1J0

12. Dundee Cement Products Ltd.
Bathurst Division
P.O. Box 170
Bathurst
E2A 3Z6

13. Edmunston Industries Ltd.
800 Canada Street
Edmunston
E3V 1W4

14. E.W. Clowater and Sons (1981) Ltd.
288 Union Street
Fredericton
E3A 1E5

15. Expert Welding Ltd.
P.O. Box 1144
Bathurst
E2A 4H9

16. Laplante Machine Shop Ltd.
P.O. Box 384
Edmunston
E3V 3L1

17. Legere Metal Works Ltd.
141 Pacific Avenue
Moncton
E1E 2G2

18. Madawaska Concrete Products
P.O. Box 250
Edmunston
E3V 3K9

19. Maritime Fiberglass Ltd.
P.O. Box 430
Minto
E0E 1J0

20. Minto Machine Welding Ltd.
P.O. Box 232
Minto
E0E 1J0

21. North York Industries Ltd.
P.O. Box 70
Nackawick
E0H 1P0

22. Ocean Steel and Construction Ltd.
P.O. Box 3187 Stn. B
Saint John
E2M 3S3

23. Quality Concrete Ltd.
648 Edinburgh Drive
Moncton
E1E 2L3

24. Rosco Steel Ltd.
P.O. Box 1148
Tracadie
E0C 2B0

25. Saint John Iron Works Ltd.
P.O. Box 7315 Stn. A
Saint John
E2L 4S7

26. Sussex Machine Shop Ltd.
P.O. Box 248
Sussex
E0E 1P0

27. Victoria Metal Works
P.O. Box 1289
Grand Falls
E0J 1M0

28. Whites Precast Concrete
RR # 4
Centreville
E0J 1H0

29. York Structural Steel Ltd.
P.O. Box 1417
Fredericton
E3B 5S4